# Bulletin
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# MECHANISM OF ERUPTION AT PARÍCUTIN VOLCANO, MEXICO

BY KONRAD B. KRAUSKOPF

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## ABSTRACT

Parícutin Volcano (Mexico) is situated on a northeast-southwest zone of weakness, which is reflected in the location of the lava vents and a satellite cone. That the lava rises from considerable depths is shown by inclusions of several rock types not known to crop out anywhere in the vicinity.

A detailed 4 month record of activity at the crater and in the area of lava outflow at the side of the cone shows little correlation between them. Only once, when the eruption became unusually vigorous, was there a direct response in the lava move-

ment. No cyclic change was observed in the eruptive behavior of the crater except a vaguely defined alternation of moderate activity and quiet in 3 week periods; no periodicity whatever appeared in the lava movement, and the amount of flowing lava remained roughly constant during the 4 months of observation.

A mechanism of eruption is suggested in which lava stands continuously at a high level in the conduit. Dissolved gas bubbles out through the crater, while the liquid spills over into a network of fissures which connect with lava vents at the side of the cone. This mechanism accounts for the simultaneous emission of huge amounts of gas from the crater and emission of lava with little gas from fissures near the base of the cone. It explains also why $SO_2$ is prominent in gas from the crater, while the small amount of gas accompanying the lava contains practically no sulfur compounds but is rich in HCl. Because the lava vents connect with the main conduit by narrow, winding channels, the lava flows are responsive only to major changes in the activity of the crater.

## INTRODUCTION

In two respects Parícutin Volcano is an especially suitable volcano for a study of the mechanism of eruption. (1) It is readily accessible: the activity of the crater can be observed continuously from a distance of less than 2 kilometers, and the cone can be frequently climbed for closer inspection; the area where lava emerges near the base of the cone, while not well situated for continuous observation, can still be visited at frequent intervals. (2) The lava is remarkably uniform in composition, so that possible complications due to differentiation or multiple lava sources do not arise. The difficulties in a study of eruptive mechanisms at Parícutin are the obvious ones which would apply to any volcano: one has only the facts of external behavior to work with; desirable facts even about external behavior are often inpossible to obtain during violent eruptions; the most impressive part of the external behavior, the emission of huge amounts of gas, cannot be quantitatively evaluated.

Speculation about the internal structure and eruptive mechanism should be based on detailed records of activity during the entire life history of the volcano. Unfortunately such records do not exist. D. E. White (1944, p. 621–627), working from personal observations and fragmentary records of the volcano's first year, traced out cycles in the eruptive activity and referred these cycles to a mechanism of accumulation and relief of gas pressure in an underground chamber, much like the mechanism which Perret (1924) deduced from observations of Vesuvius. The present work is based on 4 months of observations in the winter of 1945–1946. Since so short a time gives little information about major cycles or about the ultimate source of the lava, attention is concentrated on a few minor problems of the eruptive mechanism, on which detailed short-term observations can be expected to shed some light. The chief questions to be considered are: (1) depth of the conduit as shown by inclusions in the lava; (2) localization of the vents of gas and lava emission; (3) correlation of activity in the crater and activity in the area of lava extrusion; (4) internal structure of the cone; and (5) possible minor short-term cyclic changes.

This report is based on field work undertaken by the U. S. Geological Survey in co-operation with the Instituto de Geología de Mexico, the U. S. Committee for the Study of Parícutin Volcano, and the U. S. Department of State. (This is one of the scientific and cultural projects undertaken abroad under the auspices of the Interdepartmental Committee on Scientific and Cultural Cooperation, Department of

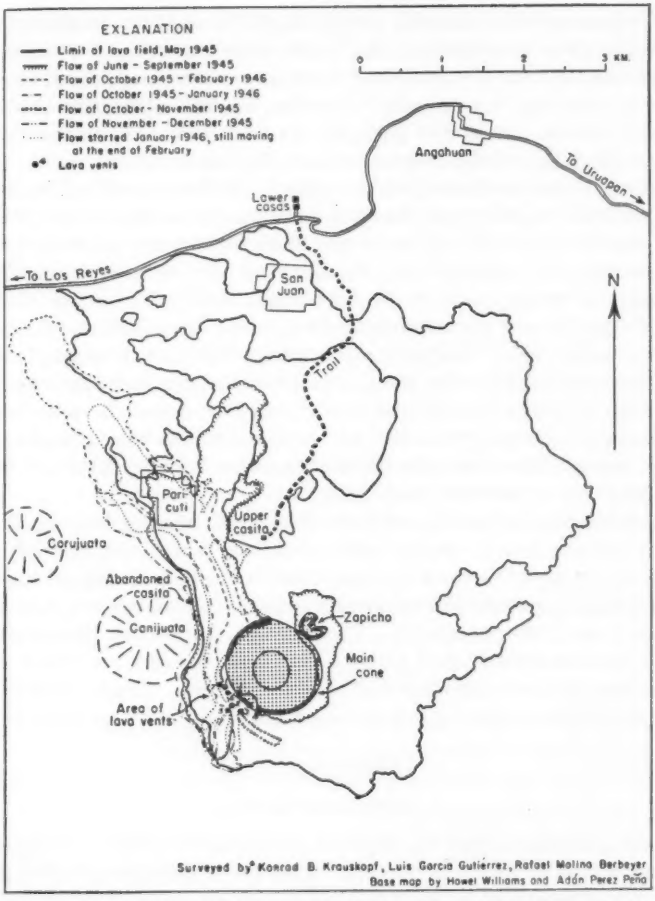


FIGURE 1.—*Sketch map of the lavas of Parícutin. June 1945 to February 1946*

State.) Facilities made available by a grant from The Geological Society of America are gratefully acknowledged. In preparing the report the writer had the advantage, of discussions with several geologists familiar with the volcano, especially E. Ordóñez Carl Fries, Jr., R. E. Fuller, L. García Gutiérrez and R. Molina Berbeyer. The writer is indebted to A. C. Waters, F. G. Wells, W. F. Foshag, and G. D. Robinson for criticism of the manuscript.

## GEOGRAPHY AND HISTORY

During the winter of 1945–1946, and for many months before, lava came exclusively from vents in the lava field near the southwest base of the cone (Fig. 1). Vents on the northeast side, which were active in the early history of the volcano, are now

covered by lava and ash. Zapicho, a small satellite cone at the northeast base, has been inactive since early 1944. Older basalt cones resembling Parícutin dot the landscape round-about. Canijuata and Corujuata are two such cones, others form the ridge on which the "upper casita" is located, and still others flank the lava field on the east. South and west of Parícutin the land rises slowly to the 12,000-foot peak of an old, deeply eroded andesite volcano, Pico de Tancítaro.

Parícutin has been continuously active since its birth in a small cultivated valley in February 1943. The general history of the volcano is merely a record of the variable but almost continuous emission of gas, steam, bombs, and ash from the crater, and a succession of lava flows from the northeast and southwest sides. This uneventful history was enlivened by the brief eruption of Zapicho in late 1943. The activity of the volcano is gradually diminishing, but at times Parícutin shows resurgences of its former vigor. The main cone has been built up to a height of about 450 meters above the original valley floor, but its lower part is so deeply inundated by lava that the highest peak now stands only some 300 meters above its immediate base. Growth was very rapid in the first few months of the volcano's existence, but the height increased no more than 12-meters between February 1945 and February 1946 (K. Segerstrom, oral communication).

Parícutin lava is monotonously uniform olivine basalt, ranging from black, aphanitic, nearly massive lava to spongy brown glass. In thin section the rock consists chiefly of tiny plagioclase laths in brown glass containing abundant black dust. Olivine appears as scattered phenocrysts up to 3 mm. in diameter and in some slides also as tiny grains in the groundmass. Pyroxene was not certainly identified but may be present in some slides as small grains. The plagioclase ($An_{55}$) is almost absent in some specimens and more abundant than the glass inothers. There is little important petrographic difference between lava in the flows and lava in bombs from the crater.

## THE CONDUIT

### DEPTH OF CONDUIT

The only information about the depth of the conduit and the nature of the rocks it cuts comes from inclusions in the lava. They are of five general types: (1) Small crystals and aggregates of quartz, fairly common in all parts of the lava. The majority are only a few millimeters in diameter, but they range up to at least 4 centimeters. The larger aggregates consist of big intergrown crystals with pronounced wavy extinction, suggesting derivation from veins or igneous rocks. (2) White spongy glass containing quartz and partly melted feldspar, found in all kinds of lava but not nearly so abundant as quartz. The origin of these inclusions is uncertain; possibly they are the result of partial fusion of granite, but Earl Ingerson (oral communication) objects that fused granite should be brown or black rather than white. (3) White or light-colored porcelanic material. Only one small fragment was found in the winter of 1945–1946; these inclusions were rare even in the earlier lavas. Milton (1945, p. 618–621) concludes from a petrographic study that the porcelanic inclusions are fragments of rhyolite or rhyolite breccia. (4) Felsic intrusive rock, ranging in texture from fine porphyritic to granitic. Some have the composition of

quartz monzonite (Schmitter, 1945, p. 113–137). These inclusions were fairly numerous in bombs early in the volcano's career, but none were found in recent lavas. (5) Anhydrite. Three crystals were found during the winter of 1945–1946, two in recent lavas and one in a bomb.

The inclusions cannot be satisfactorily identified with known rocks near Parícutin. Exposed basement rocks in the immediate vicinity are andesite and basalt, not known to be represented in the inclusions. The porcelanic rock may come from rhyolitic facies of the Tertiary lavas and pyroclastic deposits, which Williams (1945) provisionally calls the Zumpimito formation, and describes as underlying basalt and andesite near Los Reyes to the west and near Uruapan to the southeast. The anhydrite most probably comes from a sedimentary bed, possibly from gypsum associated with Lower Cretaceous strata which are known to crop out near Lake Chapala about 100 kilometers northwest (E. Ordóñez, oral communication). The difficulty with this suggestion is that no other inclusions of definitely sedimentary origin have been found; conceivably some of the quartz inclusions are recrystallized quartzite, but nothing in their texture suggests this origin. The felsic intrusive rocks cannot be matched by any rocks known to crop out within at least 50 kilometers of Parícutin. If all the quartz and glass fragments are derived from felsic intrusive rocks, the large quantity of these inclusions erupted during the history of the volcano must mean that the conduit tranverses a great thickness of such rocks.

In summary, inclusions of several rock types not known to crop out near the volcano indicate that the lava rises from considerable depth, but they give no further information about the conduit.

### LOCALIZATION OF VENTS

According to reports of eye witnesses, the small cone formed in the earliest days of eruption was elliptical, elongated northeast-southwest, but rapidly became circular. The influence of a northeast-southwest plane of weakness was shown later by the emission of lava flows exclusively from the northeast and southwest sides of the cone, by the elongation of the inner crater in this direction when more than one vent was active, and by the appearance of Zapicho at the northeast base.

Why should the eruption be restricted to a pipelike conduit, instead of spreading along the controlling northeast-southwest fissure? No observer of the early stages has reported a cross fissure which might help to localize the conduit, and there is no suggestion of one in the later activity. If the evolution of gas were quiet the localization might be simply a chimney effect, the gas seeking always the highest point of the growing mound because this point provides the longest column of heated vapor. But the gas is normally emitted so forcefully that the chimney effect can hardly play an important role in determining the shape or position of its orifice. Perhaps the restriction to a circular pipe is the result of erosion by the rapidly ascending gas (Hadley Bramel, oral communication). If one part of the original fissure happens to be even a littler wider than adjacent parts the gas flow and the consequent erosion will be greater there, so that this wider part will increase in size until practically all of the gas is concentrated in it.

The general areas in which fluid lava comes to the surface are localized by the northeast-southwest zone of weakness, but the precise locations of the vents or "bocas" of individual flows are determined by local factors.[1]

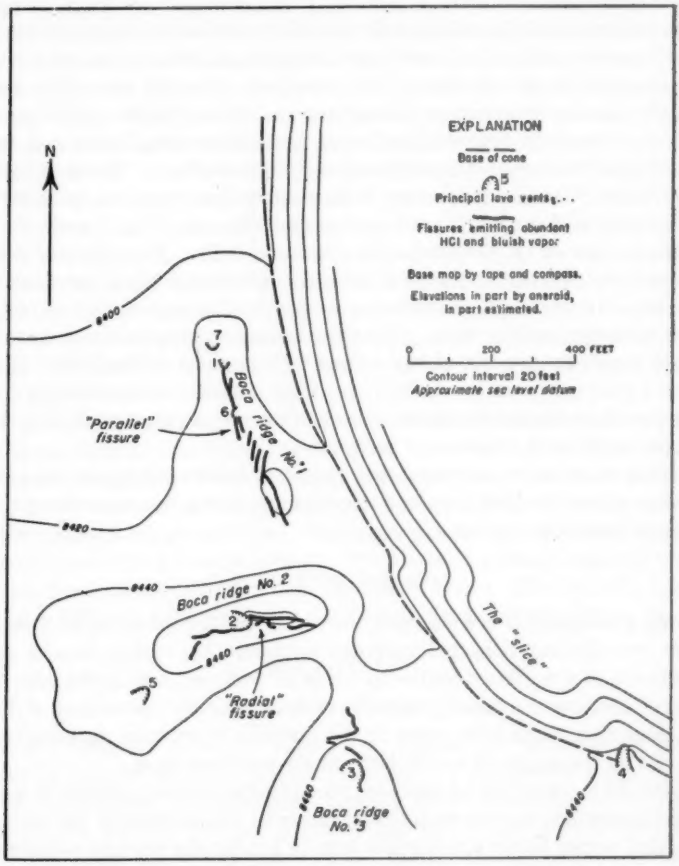


FIGURE 2.—*Area of lava vents. Rough contour map, November 1945*

All the flows observed by the writer originated in an area measuring about 300 by 700 meters on the lava surface at the southwest base of the cone. During the winter of 1945–1946 nearly all the flowing lava came from vents along two conspicuous fissure zones, Boca ridge No. 1 and Boca ridge No. 2 (Fig. 2). One flow in early November cam from Boca ridge No. 3, and in early March another broke out from approximately the site of the old Boca No. 4 (also called locally the "Aguan vent") at the

[1] Boca, Spanish for "mouth", is the common term used at Parícutin for a vent from which a lava stream is flowing or has recently flowed. It is used in this sense in Figures 2–8.

base of the cone. The fissures on Boca ridges 1 and 2 are discontinuous, branching cracks in the old lavas, conspicuous because of the bluish vapor continually rising from them and the light-colored sublimates deposited on their edges. The vents

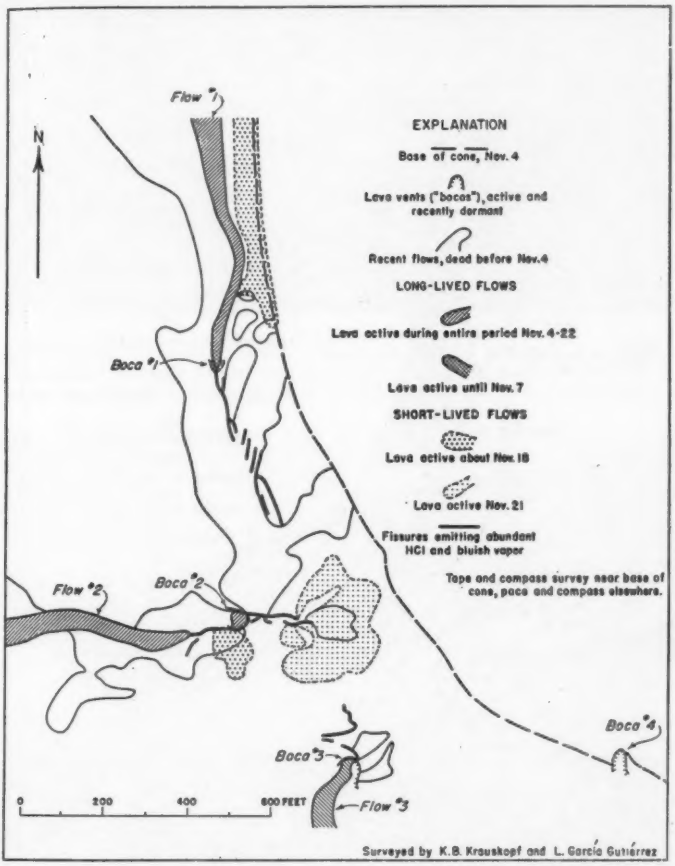


FIGURE 3.—*Area of lava vents. November 4 to November 22, 1945*

themselves are small sections of the fissure zones which change position from time to time. A vent may remain in the same spot for several weeks, it may migrate along the fissure zone, or it may cease flowing and be supplanted by another vent. The sites of the principal vents active between October 1945 and March 1946 are indicated by numbers on Figure 2; the important changes in the vents and in the lava streams from them are shown by the series of maps in Figures 3 to 7.

The principal fissure zones (Fig. 2 and Pl. 1, fig. 1) lie along the crests of ridges. This is probably not because the fissures prefer ridgetops, but because the ridges are built up by minor flows of lava from various points along the fissures, independ-

ent of the principal vents. More difficult to understand is the common occurrence of an individual vent at the highest point along its fissure ridge. In many cases, however, the vent may only seem to stand at the highest point. The lava may actu-

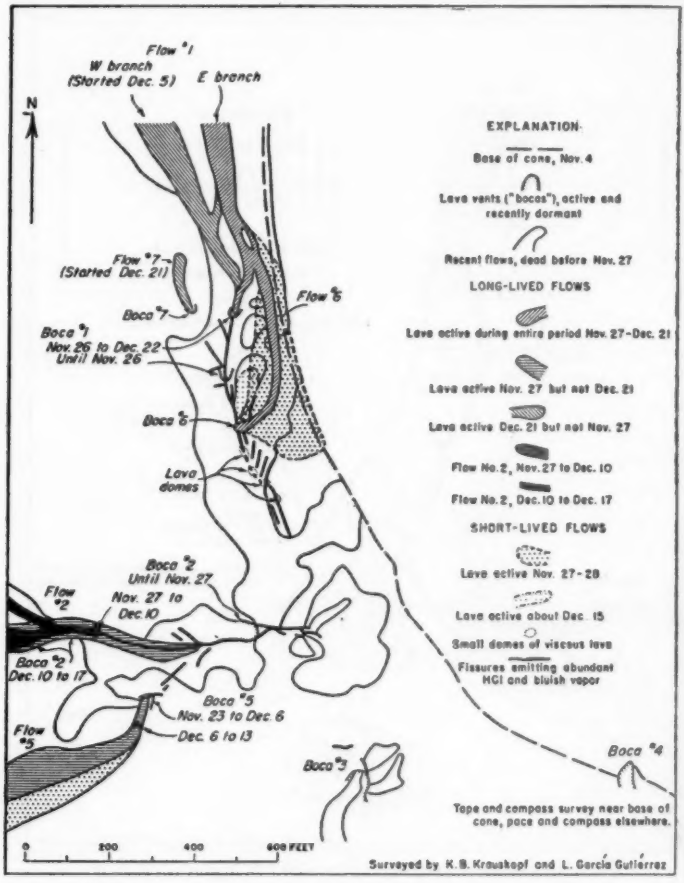
Flow °1
W branch
(Started Dec. 5)
E branch
N
Flow °7
(Started Dec. 21)
Boca °7
Flow °6
Boca °1
Nov. 26 to Dec. 22
Until Nov. 26
Boca °6
Lava domes
Boca °2
Until Nov. 27
Flow °2
Nov. 27 to Dec. 10
Boca °2
Dec. 10 to 17
Boca °5
Nov. 23 to Dec. 6
Dec. 6 to 13
Flow °5
Boca °3
Boca °4
EXPLANATION
Base of cone, Nov. 4
Lava vents ("bocas"), active and recently dormant
Recent flows, dead before Nov. 27
LONG-LIVED FLOWS
Lava active during entire period Nov. 27-Dec. 21
Lava active Nov. 27 but not Dec. 21
Lava active Dec. 21 but not Nov. 27
Flow No. 2, Nov. 27 to Dec. 10
Flow No. 2, Dec. 10 to Dec. 17
SHORT-LIVED FLOWS
Lava active Nov. 27-28
Lava active about Dec. 15
Small domes of viscous lava
Fissures emitting abundant HCl and bluish vapor
Tape and compass survey near base of cone, pace and compass elsewhere.
0 200 400 600 FEET
Surveyed by K. B. Krauskopf and L. García Gutiérrez

FIGURE 4.—*Area of lava vents. November 27 to December 21, 1945*

ally emerge from a considerable length of fissure extending from the summit down the side or end of the ridge. Part of the lava rises to the top of the ridge because the fissure is too narrow to accommodate the entire flow in its lower part. This mechanism is suggested by the fact that, when a flow starts, its vent may migrate *up* the ridge (*note* Boca 7, Fig. 5), while at the death of a flow the vent commonly migrates *down* the ridge (*note* Bocas 2 and 5, Fig. 4; Boca 1, Fig. 5). The vents of short-lived, gas-rich flows (for instance, the November 21 flow on Fig. 3) are

FIGURE 1. BOCA RIDGE NO. 1

Looking north from the east end of Boca ridge No. 2. Base of cone on right; Canijuata on left. In early November, when the picture was taken, only a single flow was moving from this ridge; its vent is at the far end, where the white vapor is most dense. Photograph by L. García Gutiérrez.

FIGURE 2. SOUTH SIDE OF CONE IN MARCH 1946

The "slice" is at the left, with a new flow issuing from Boca No. 4 at its eastern base. The scar on the side of the cone was caused by partial collapse when the new flow began. Photograph by courtesy of E. Ordóñez.

## LAVA VENTS

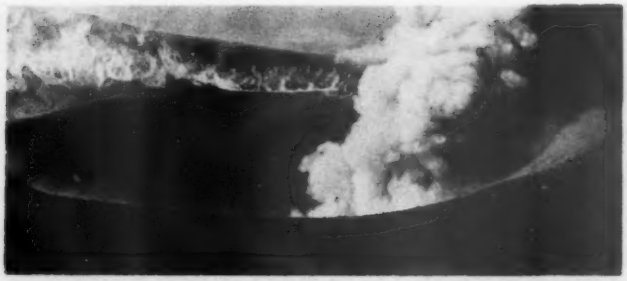

FIGURE 1. SMALL INNER CONE OF FRAGMENTAL MATERIAL WITHIN THE MAIN CRATER

Typical of the periods of moderate activity in early November and late December. In left background steam rises from cracks in the outer part of the crater. Photograph by L. García Gutiérrez.

FIGURE 2. INNER CRATER IN MID-DECEMBER

When several vents were active simultaneously. Picture taken from edge of outer shelf looking southwest. Small crater in the foreground (about 12-meter diameter) contains at least three small vents a few meters apart, all emitting vapor forcefully but the one on the right more active than the others. In left background, about 50-meters away, a darker but more tenuous cloud rises slowly from another small crater. Photograph by L. García Gutiérrez.

FIGURE 3. THE THROAT OF THE INNER CRATER

At a time when only transparent gas was rising from it. Pyroclastic material loose and roughly stratified near top, partly welded and incandescent at bottom. Photograph by L. García Gutiérrez.

INTERIOR OF THE CRATER

probably localized by a different process. Wherever lava rises from below into the bottom of an open fissure, spattering and overflow quickly build a conduit to the top of the fissure, thus localizing the vent at that point.

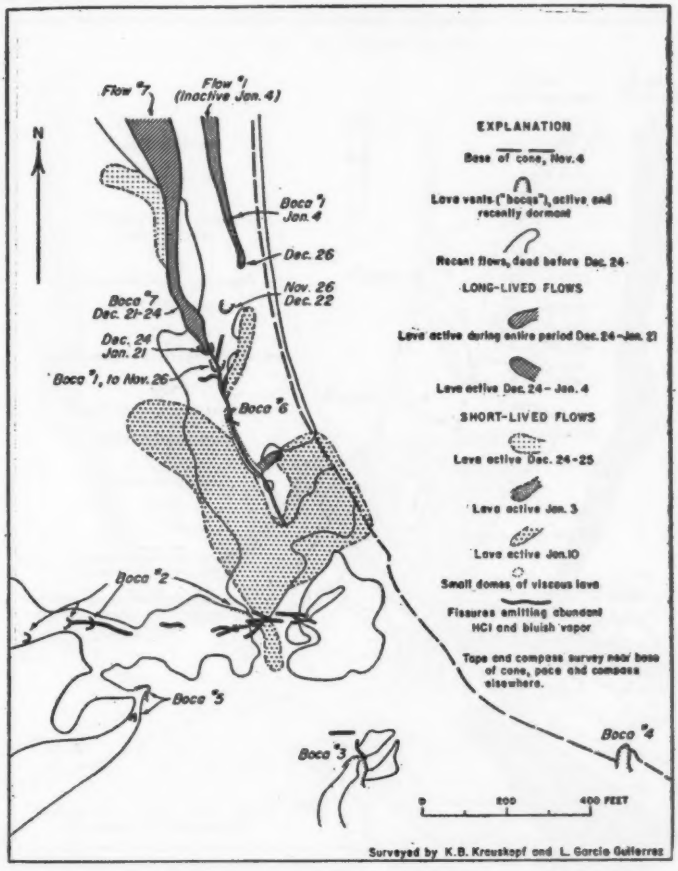
Flow "7
Flow "1 (inactive Jan. 4)
N
Boca "1 Jan. 4
Dec. 26
Nov. 26 Dec. 22
Boca "7 Dec. 21-24
Dec. 24 Jan. 21
Boca "1, to Nov. 26
Boca "6
Boca "2
Boca "5
Boca "3
Boca "4
EXPLANATION
Base of cone, Nov. 4
Lava vents ("bocas"), active and recently dormant
Recent flows, dead before Dec. 24
LONG-LIVED FLOWS
Lava active during entire period Dec. 24–Jan. 21
Lava active Dec. 24 – Jan. 4
SHORT-LIVED FLOWS
Lava active Dec. 24-25
Lava active Jan. 3
Lava active Jan. 10
Small domes of viscous lava
Fissures emitting abundant HCl and bluish vapor
Tape and compass survey near base of cone, pace and compass elsewhere.
0 200 400 FEET
Surveyed by K.B. Krauskopf and L. Garcia Gutierrez

FIGURE 5.—*Area of lava vents. December 24, 1945 to January 21, 1946*

The orientation of the two principal fissure zones, one parallel to the base of the cone and one approximately radial to it, suggests, at first, that they might have some simple relation to the structure of the cone—one being the fissure of a ring-dike or cone-sheet, the other the fissure of a radial dike. Nothing in the activity along the two zones, however, bears out this suggestion. As will be shown, the fissures are probably not of deep origin but rather are the places where lava finally reaches the surface after working its way out from under the cone through cracks in older flows.

## CORRELATION OF DIFFERENT KINDS OF ERUPTIVE ACTIVITY

### PROCEDURE

A detailed record of activity in the area of lava vents and in the crater was kept for the 4-month period of observation, in the hope that the different kinds of activity

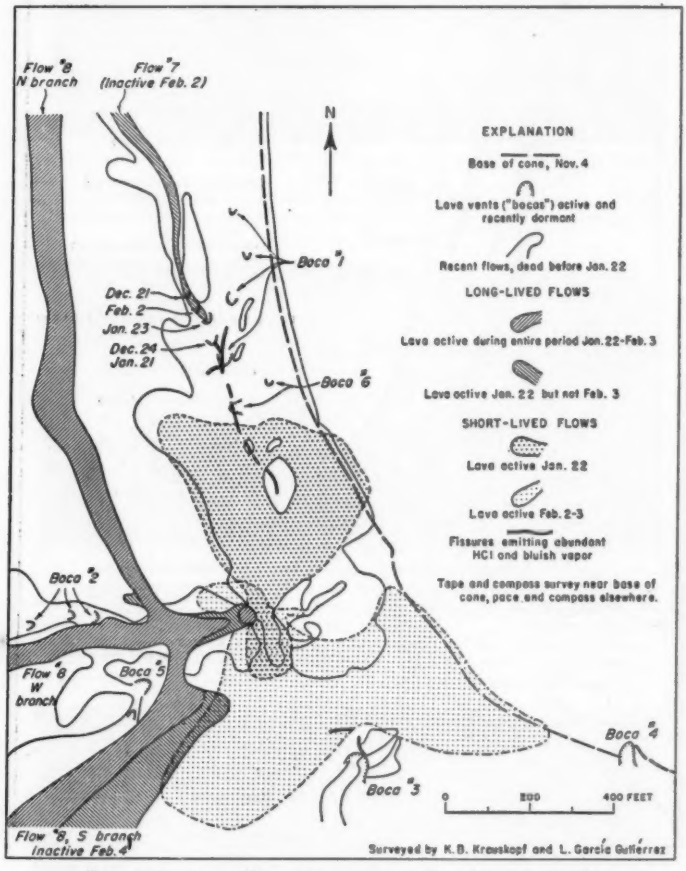


FIGURE 6.—*Area of lava vents. January 22 to February 3, 1946*

could be correlated. To obtain the record visits were made to the lava vents every 2 to 4 days and all significant changes were noted on maps (Figs. 3–7); the activity of the crater as seen from below was recorded daily, the cone was climbed once a week (eruptive activity permitting), and changes in the shape of the crater were sketched (Figs. 8–12).

### FLOW ACTIVITY

In the area of lava bocas three kinds of lava movement are conspicuous: (1) long-lived flows which move for at least a week and attain lengths of at least half a kilo-

meter, generally welling up quietly from their bocas; (2) short-lived flows which are in motion for less than 2 days and attain lengths of less than 200 meters, generally emerging from their bocas accompanied by much gas and spattering of lava; and (3)

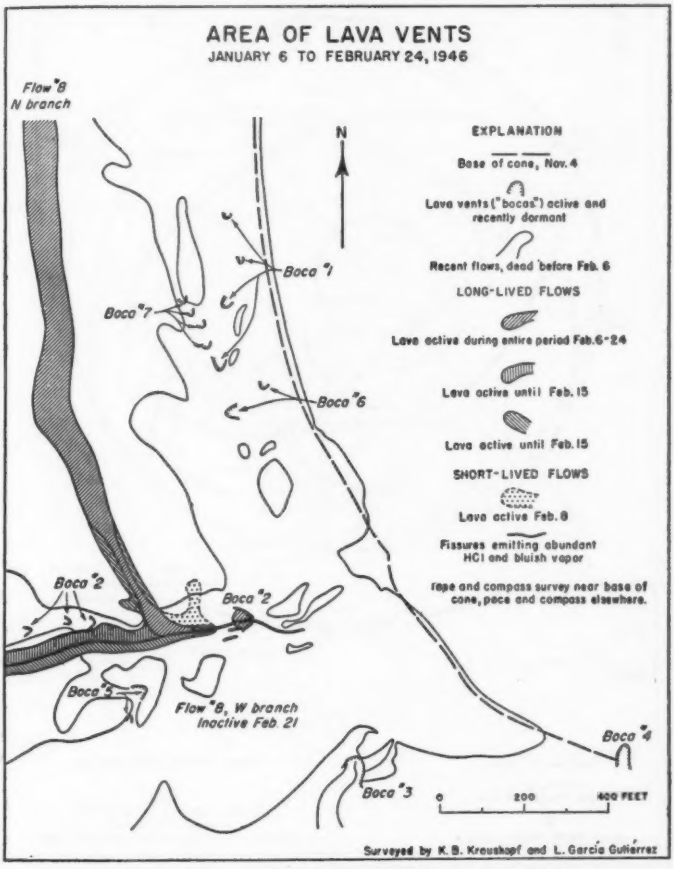


FIGURE 7.—*Area of lava vents. February 6 to 24, 1946*

the slow movement of viscous lava out of fissures to form elongated domes and thick curved sheets. All types of lava activity at the bocas are accompanied by emission of gas, of which the principal odor is that of hydrogen chloride; rough chemical tests for the sulfur gases showed no more than traces. The amount of lava flowing from the bocas, as indicated by calculations on rates of lava movement, probably was seldom less than 200,000 or more than 500,000 cubic meters per day during the 4-months of observation.

No simple relations are evident among lava movements from the different fissure zones. The lava shows the same apparent disregard for differences in hydrostatic

pressure that has been commented on at other volcanoes. Two flows can move rapidly and persist for weeks when one has a boca 40 or 50 meters higher than the other. In early November three flows were moving, but No. 3 was sluggish and soon

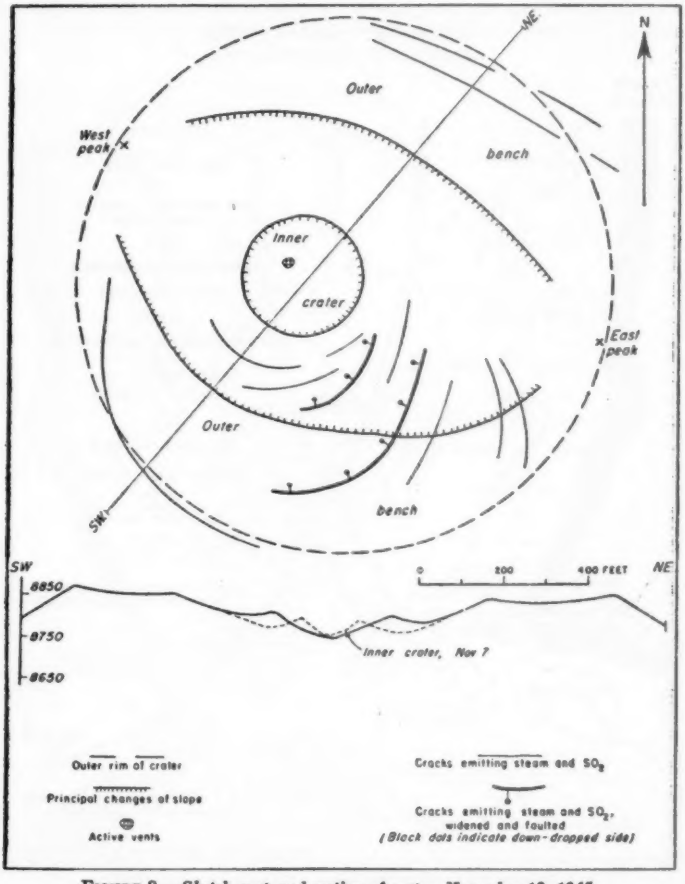


FIGURE 8.—*Sketch-map and section of crater, November 13, 1945*

died (Fig. 3). In late November two new vents opened up; although flow No. 2 began to die at this time, probably more lava was in motion between November 27 and December 6 than at any other time (Fig. 4). All activity on Boca ridge 2 ceased in mid-December and, except for one minor flow on January 10, was not resumed until January 22 (Fig. 5). On this date a spectacular eruption from the old Boca 2 sent lava cascading down the ridge in several directions; the movement presently settled down into a three-branched flow (Fig. 6). Meanwhile the flow from Boca ridge 1 began to lessen and by early February had died completely. The new flow from Boca

2 remained active during the month, but two of the original three branches became stagnant (Fig. 7). Thus during November and early December lava moved from both of the principal fissure zones, but thereafter from only one at a time.

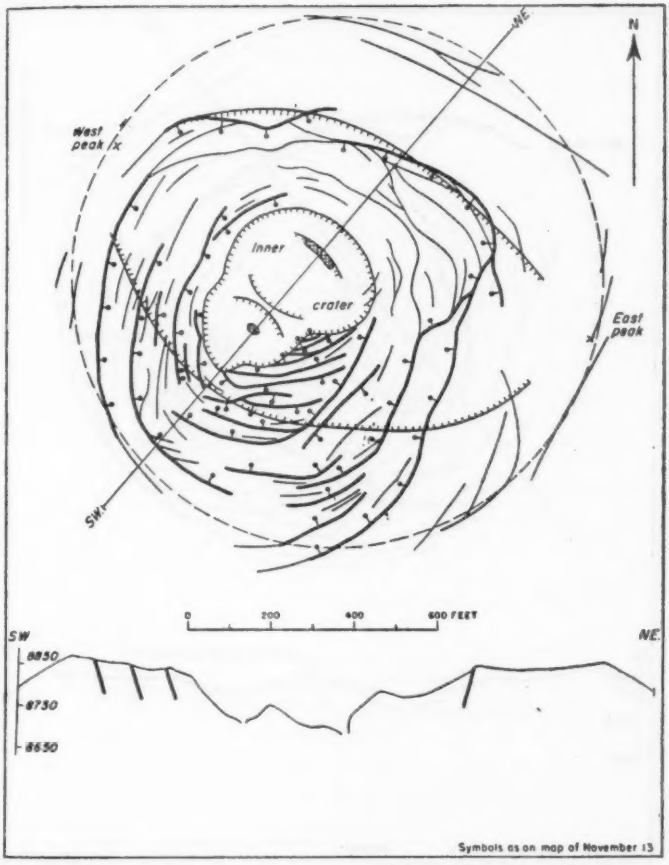


FIGURE 9.—*Sketch-map and section of crater, December 4, 1945*

An important change in the boca area occurred in early March, after the writer had left the volcano. The segment of the cone marked "the slice" in Figure 2, originally pushed out during the birth of a flow from Boca No. 4 in December 1944, had been moving very slowly outward and tipping through a steeper and steeper angle all during the winter months. Finally in March the segment was pushed still farther out, the south wall of the cone partly collapsed, and a new flow appeared approximately at the site of the old Boca No. 4 (Pl. 1, fig. 2). As this flow started, movement from Boca No. 2 quickly ceased.

## EXPLOSIVE ACTIVITY

Activity in the crater consists of the emission of gas, ash, solid lava blocks, and bombs of fluid lava. Relative and total amounts of these materials vary greatly

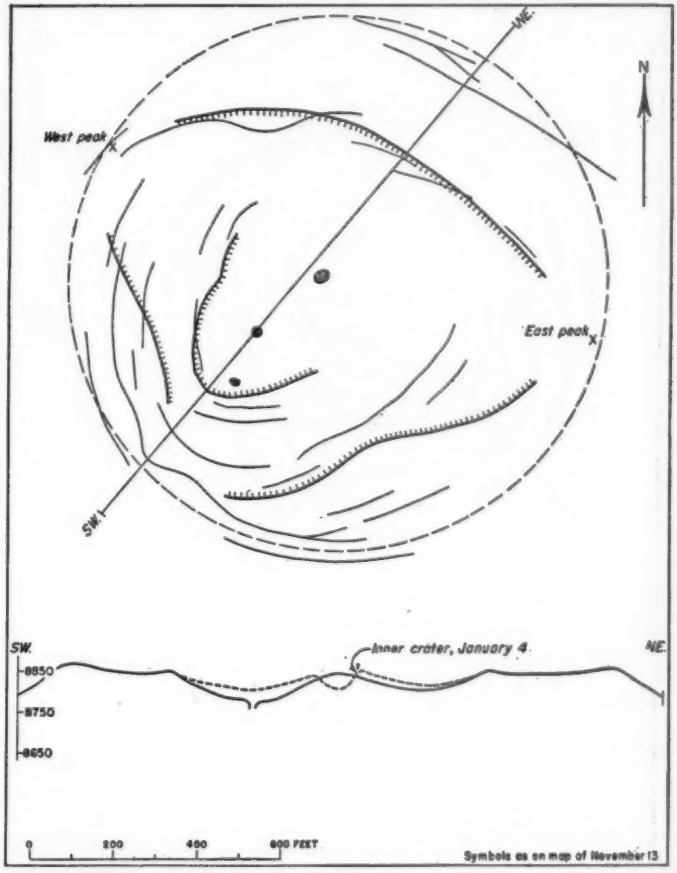


FIGURE 10.—*Sketch-map and section of crater, December 18, 1945*

from time to time; the manner of gas emission changes from quiet to explosive; and sound effects go through an extensive repertoire of pounding, rumbling, whistling, puffing, and complete silence. No method of collecting or analyzing gas from the crater suggested itself, but certainly the gas contains abundant water vapor, which condenses in the eruption cloud, and enough sulfur dioxide to make the crater at times uninhabitable. That sulfur is one of the gases rising through fissures in the crater is shown by abundant needles of monoclinic sulfur deposited at the mouths of the fissures.

Although changes in the activity of the crater are discouragingly erratic, some general types of behavior correlate with alterations in the crater's shape. In late October and early November the cone was fairly active, with a dense white eruption

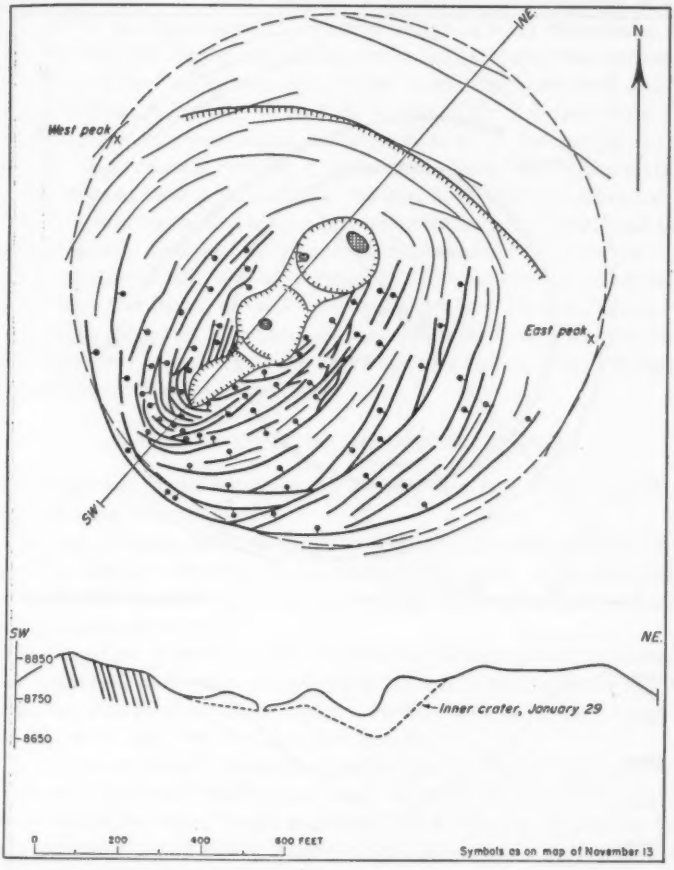


FIGURE 11.—*Sketch-map and section of crater, January 19, 1946*

cloud most of the time and frequent showers of bombs. During this period the crater consisted of two very shallow nested basins, with a small cone in the inner one built of bombs and blocks dropped around the orifice (Fig. 8 and Pl. 2, fig. 1). From mid-November until December 7 the eruption was less showy, the cloud for the most part consisting of a narrow, tenuous, gray column and no bomb showers rising over the edge of the crater. During this period the inner cone vanished, the interior of the crater deepened, a second vent was formed, and an elaborate pattern of cracks and small faults developed around the inner crater (Fig. 9). During the last 3

weeks of December eruption was more active the crater again filled with debris, and a small inner cone was formed (Fig. 10). Another quiet period followed in early January, again producing a deeper crater and a pattern of concentric faults (Fig. 11).

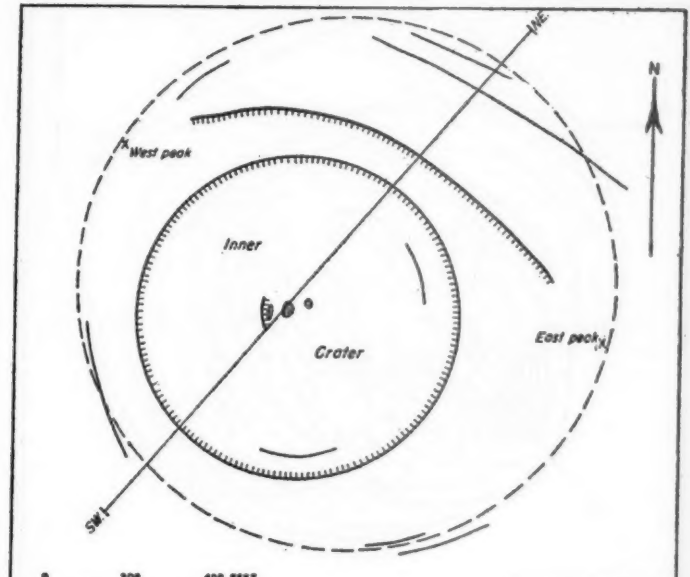


FIGURE 12.—*Sketch-map and section of crater, February 12, 1946*

When violent activity recommenced on January 21 the pattern changed: The eruption cloud became thick and dark, sprinkling ash liberally over the surroundings, and bomb showers were more frequent and more violent than at any previous time. The crater developed a huge inner funnel, and the pattern of concentric cracks disappeared (Fig. 12). This type of activity continued all through February. What changes in the crater accompanied the formation of the new flow in early March are not known to the writer.

It is probably significant that with the most pronounced change in the activity of the crater, the change from a quiet period to a period of exceptionally violent eruption on January 21 and 22, the lava flow from Boca ridge 1 stopped, and a big new flow began from Boca ridge 2. One is tempted to speculate that an upsurge of lava into the lower part of the cone caused the whole structure to swell, thus closing a fissure parallel to the base of the cone and opening one radial to it; and that continued pressure of the lava on the interior caused the "slice" to move outward until finally a new flow broke out through its base, sapping the supply of the flow from Boca 2. Such speculations are not very convincing, however, since up to January 21 there seems to be no hint of correlation between activity at the bocas and in the crater. The times of greatest and most rapid change in the lava flows, late November and December 21–24, were unaccompanied by any unusual activity in the crater; and the bocas were not appreciably disturbed by changes in eruptive activity in the crater in mid-November, early December, and early January. Nor did the sudden appearance of short-lived but active bocas, as on November 21, January 10, and February 2, lead to any change in the crater. Perhaps one may generalize that activity in the crater is fairly independent of activity at the bocas but that certain major changes in one are accompanied by changes in the other.

## INFERRED INTERNAL STRUCTURE OF AREA

A satisfactory hypothesis regarding the internal structure of the cone should explain the following observations:

(1) The cone, although built up almost entirely of fragmental material, is able to withstand the explosive emission of gas from a small vent. The size, shape, and position of the vent change frequently, but even the most violent bursts of gas do not produce major changes suddenly.

(2) Emission of gas is almost entirely confined to the central vent, emission of lava to the boca area. The gas that accompanies the lava is negligible compared to the huge quantities that pour out of the crater. Furthermore, gas from the central vent is rich in $SO_2$, while that from the bocas is rich in HCl.

(3) The extensive inward slumping of the crater at times of quiet eruption, the slumped material evidently settling into space below occupied by fluid, and the outward movement of the "slice" on the southwest side of the cone, apparently produced by pressure of gas or liquid within, suggest that a considerable body of gas or liquid is present within the cone.

(4) At times very numerous large shapeless chunks of incandescent liquid are blown out of a small vent in the crater, suggesting that liquid lava must fill the central conduit nearly to its top.

(5) Activity at the crater is related to activity in the boca area in that certain major changes in one are reflected by changes in the other, but there is no appreciable correlation with respect to minor changes.

(6) Activities in adjacent bocas show only slight interrelationships.

(7) When two or more vents appear in the crater, their activity is usually somewhat different. Gas may shoot out of one with explosive violence, but rise in slow

puffs from the other; both may show explosive activity, but with different timing; one may throw out showers of molten, incandescent fragments, while the other emits

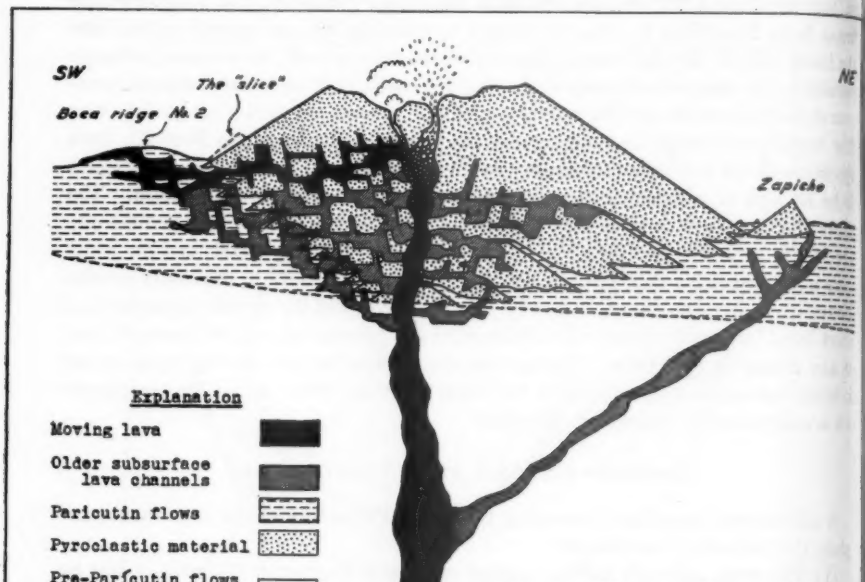


FIGURE 13.—*Inferred vertical NE-SW section through Paricutin, December 1945.*

only a little fine dust (Pl. 2, fig. 2). This independence in activity of closely adjacent vents has been observed at other volcanoes (Washington, 1917).

A possible structure to account for these facts is shown in Figure 13, a rough vertical section along a northeast-southwest line through Boca ridge 2 and Zapicho. The "slice", not actually in the section, is shown by a dashed line. The floor of the volcanic structure, indicated by a dashed line, is guess work, based simply on the fact that the original surface sloped generally upward toward the southwest. Because the cone was built up of fragmental material while successive lavas inundated its base, the lower edge of the fragmental pile is shown schematically as an interfingering of ash and lava. The central conduit is approximately vertical within the cone, but its position in the basement rocks can only be guessed.

The ability of the cone to withstand violent emission of gas from a small opening suggests that the interior cannot consist wholly of loose debris. Presumably near the lava conduit or conduits the fragmental material is partly welded into a fairly solid rock. This is indicated also by observations in the throat of the crater when an exceptionally quiet eruption permits close inspection. The walls of the inner crater

are made up of roughly stratified angular debris which grades downward into massive incandescent rock (Pl. 2, fig. 3). On the one occasion when a temperature measurement was possible this incandescent material gave a maximum reading of 980°C., about 20° cooler than the lowest temperature recorded for flowing lava. Some of the solid blocks thrown out of the crater are composed of angular fragments so firmly cemented that their outlines are in part indistinct.

The welding is not complete enough, however, to give the interior of the cone great strength. New vents open in the crater at frequent intervals, showing that gas can force its way upward through new channels when old ones become clogged; and the slumping in the crater during times of quiet eruption must mean that the material of the cone can be easily cracked.

Lava as shown on the diagram nearly fills the central conduit and spills over into the vent area through narrow, branching fissures. Within the conduit the lava must be in continual violent agitation as gas comes out of solution and bubbles up through it. Fragments of lava splash upward as the bubbles burst, fall back, and again are carried up. The continual up-and-down motion, accompanied by the tearing apart of liquid fragments and their constant exposure to streaming gas, removes the less soluble gases (notably the sulfur gases) so that the liquid which ultimately finds its way to the lava vents contains only small amounts of the more soluble gases (notably HCl). As the boiling proceeds in the conduit, fresh lava with dissolved gas is continually supplied from below.[2]

The channels connecting the conduit with the area of lava vents are shown as a network of fissures, partly radial and partly concentric around the conduit. The channels must be sufficiently narrow and winding so that temporary changes in the amount of lava supplied to the conduit or in its gas content do not immediately affect lava movement into the vents. If the connection from one lava vent to another is through similar tortuous channels, the independence in vent activity and the apparent disregard of the vents for hydrostatic equilibrium are understandable. Perhaps a similar branching and irregularity of channels may even explain the existence of two or more gas vents in the crater with different kinds of eruptive behavior: If, for instance, a side channel which is connected to the main conduit by a small opening has a wide area for an expansion chamber, then explosive activity in the main channel could be accompanied by quiet evolution of gas from the other vent. Or one might imagine the entrance to a side channel becoming temporarily clogged by splashing lava, so that its activity would be timed differently from that in the main channel.

The lava channels change from time to time, probably in part because of erosion by moving lava, in part by slumping, and in part because of pressure exerted by lava and gas within the cone. But in general they remain close to the northeast-southwest zone of weakness which initially determined their position.

The observed slumping in the crater at times of quiet eruption may be due either to settling into an enlarged portion of the central conduit or to collapse of some of the channels which connect the conduit with the bocas. The slow outward movement of the "slice" is explained by the push against it of lava in fissures near the side

[2] In general outline this mechanism is similar to the one suggested by Perret (1924, p. 75) for Vesuvius.

of the cone; in early March the pressure was sufficient to break through the side of the cone and form a new flow.

This mechanism, unlike others suggested (White, 1945) assumes that lava destined for the boca area rises high in the central conduit and overflows, instead of branching from the main stream at depth. Assuming a branch at low levels makes it difficult to account for the separation of gas and liquid. Why should the material rising on one branch be greatly different from that in the other? Perhaps the formation of the satellite cone Zapicho, in which the eruptive activity was similar to that in the main crater, represents a branching of the conduit at depth, but the slow welling up of lava in the boca area requires a different explanation.

## CYCLIC CHANGES

A cyclic alternation of abundant flowing lava from the bocas and vigorous gas emission from the crater, similar to the alternation described by White in the first year of the volcano's history, is suggested by observations of Howel Williams in the winter and spring of 1945 (Krauskopf and Williams, 1946). But during the winter of 1945–1946 there was little sign of such cyclic alternation. The crater's increased eruptive activity in late January and February followed by a big new lava flow from the base of the cone in March resembles the sequence of events in White's cycle, but there was no diminution in lava activity during the violent eruptive period as required by the cycle. Probably the volcano has reached a stage of activity in which White's cycle is no longer prominent—perhaps because the average gas content of the lava is now smaller, so that "boiling" takes place high in the conduit rather than at lower levels.

The only indication of cyclic repetition of events in this period of observation (1945–46) was the change in activity of the crater from moderately active to quiet and the reverse at approximately 3-week intervals between late October and late January. These alternations were unaccompanied by any significant changes in the emission of lava. They seem more plausibly explained by slight differences in the average level of lava in the conduit, or differences in its gas content, than by the periodic blowing off of gas from a magma chamber at depth.

The day-to-day behavior of the volcano, far from showing any sign of short-term periodicity, remained completely unpredictable. Intervals of quiet, periods of steady gas emission, periods of dull explositions followed one another with complete irregularity. Sometimes a single loud burst of gas would interrupt a period of quiet which otherwise endured for several hours. The form of the eruption cloud sometimes remained almost the same for days, sometimes changed from hour to hour. The three-week periods labeled "quiet" and "moderately active" are rough generalizations only; within each period activity appropriate to the other type was not uncommon. And during the entire 4 months of observation the amount of flowing lava remained surprisingly constant.

The continual erratic minor variations in activity, combined with the virtual absence of either long-term or short-term cycles and the almost constant outflow of lava, suggest that during the winter of 1945–1946 lava remained high in the conduit of the

volcano but was subject to slight variations in the supply from below. The greater activity of late January and February perhaps means an increase in the amount of lava or its gas content or both. There is no evidence for separation of lava and gas in a magma chamber at depth, but on the contrary the great difference in the amount and kind of gas emitted from the crater and from the lava bocas seems good evidence that the separation takes place largely within the cone itself.

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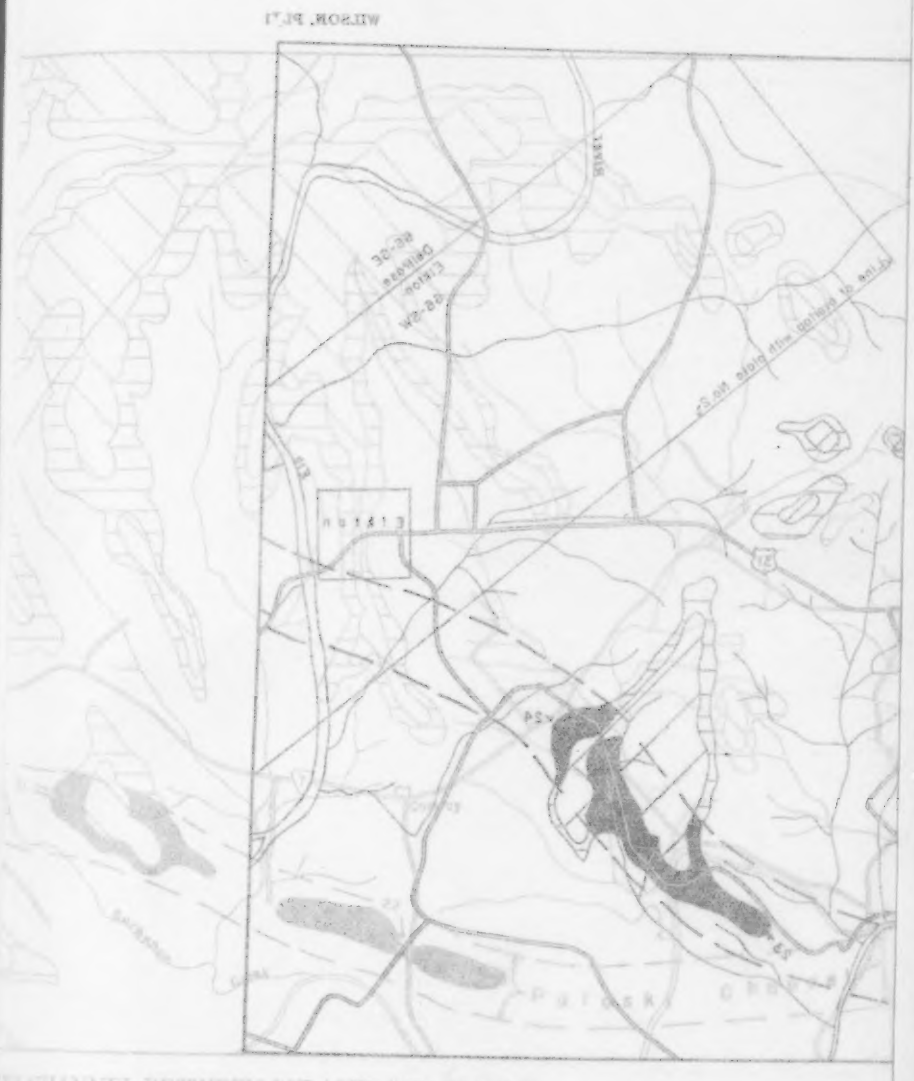

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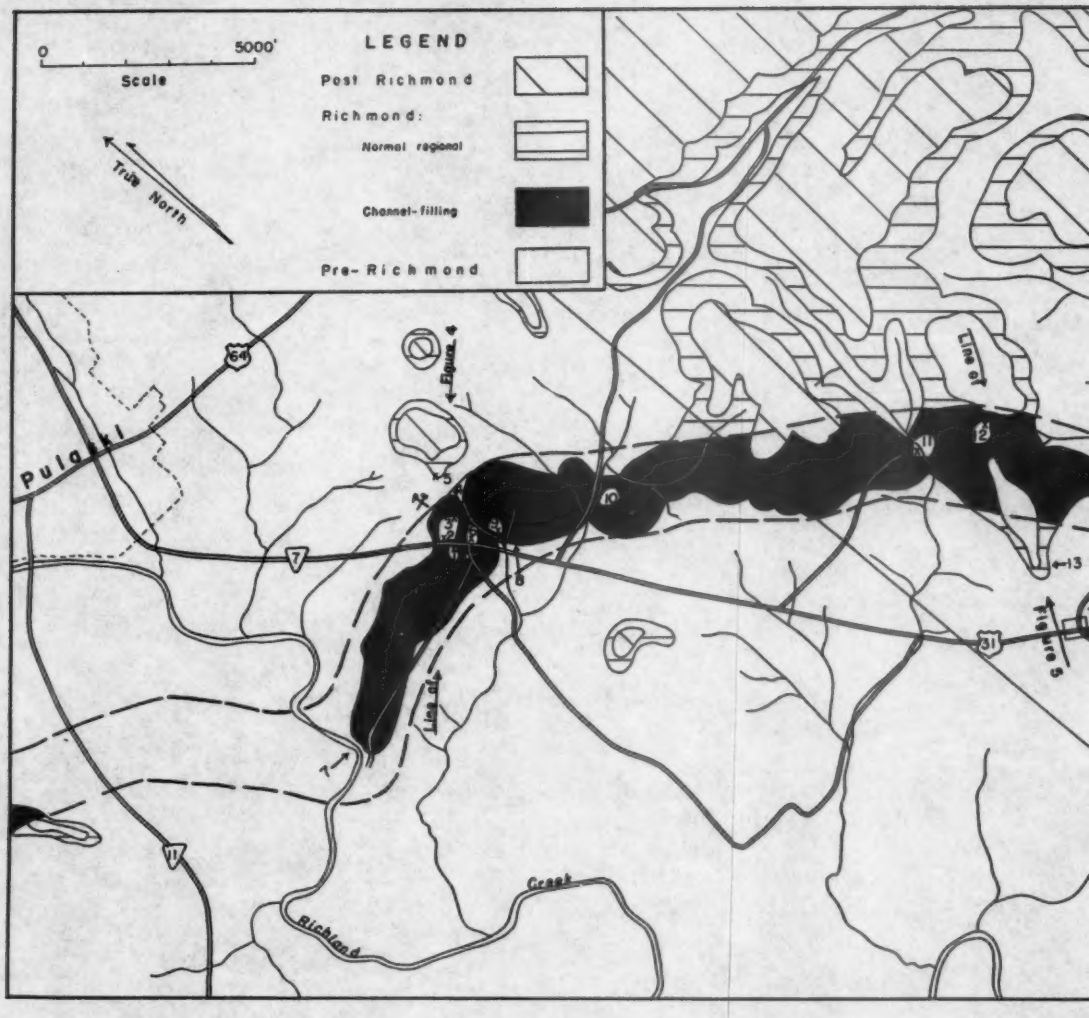
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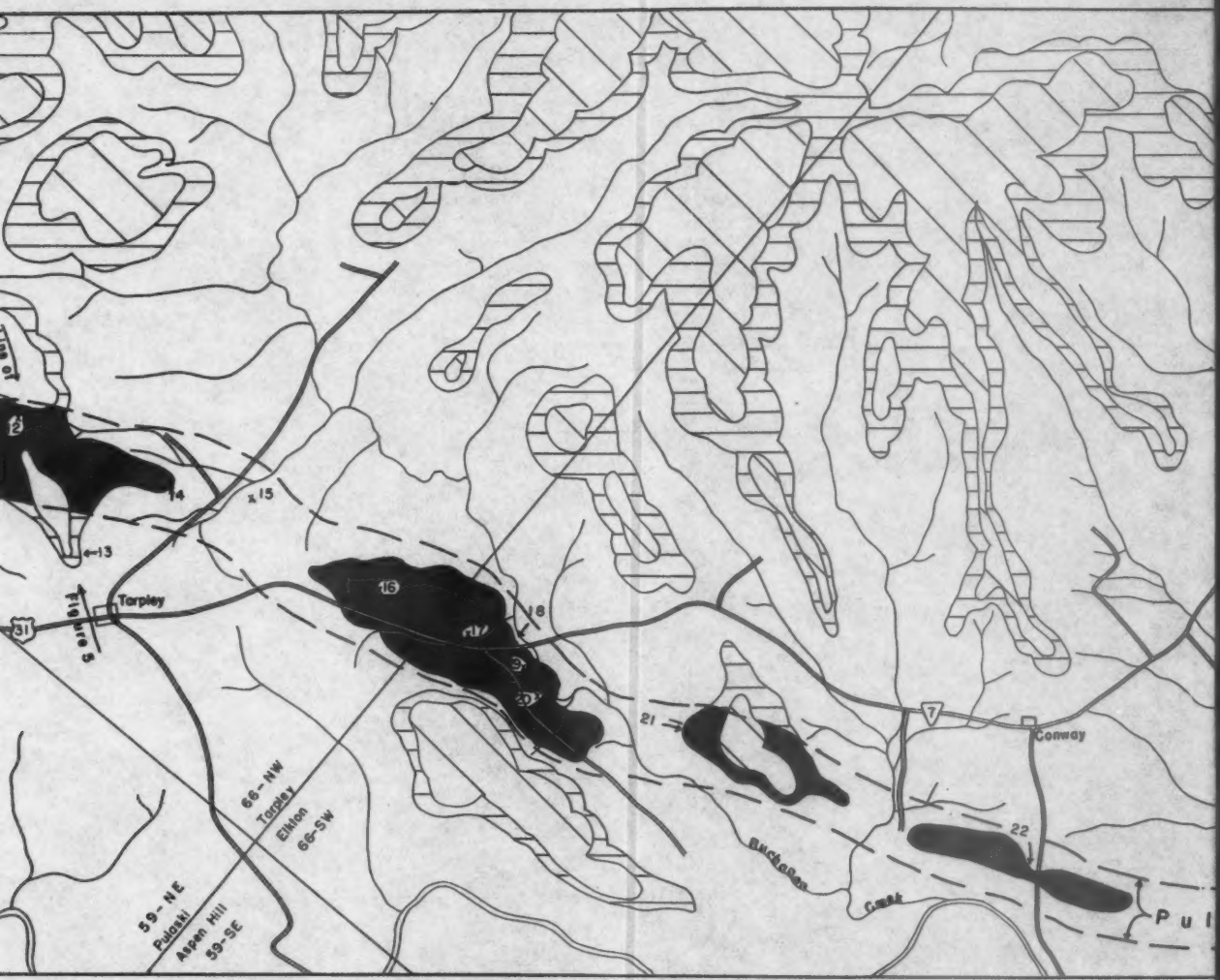
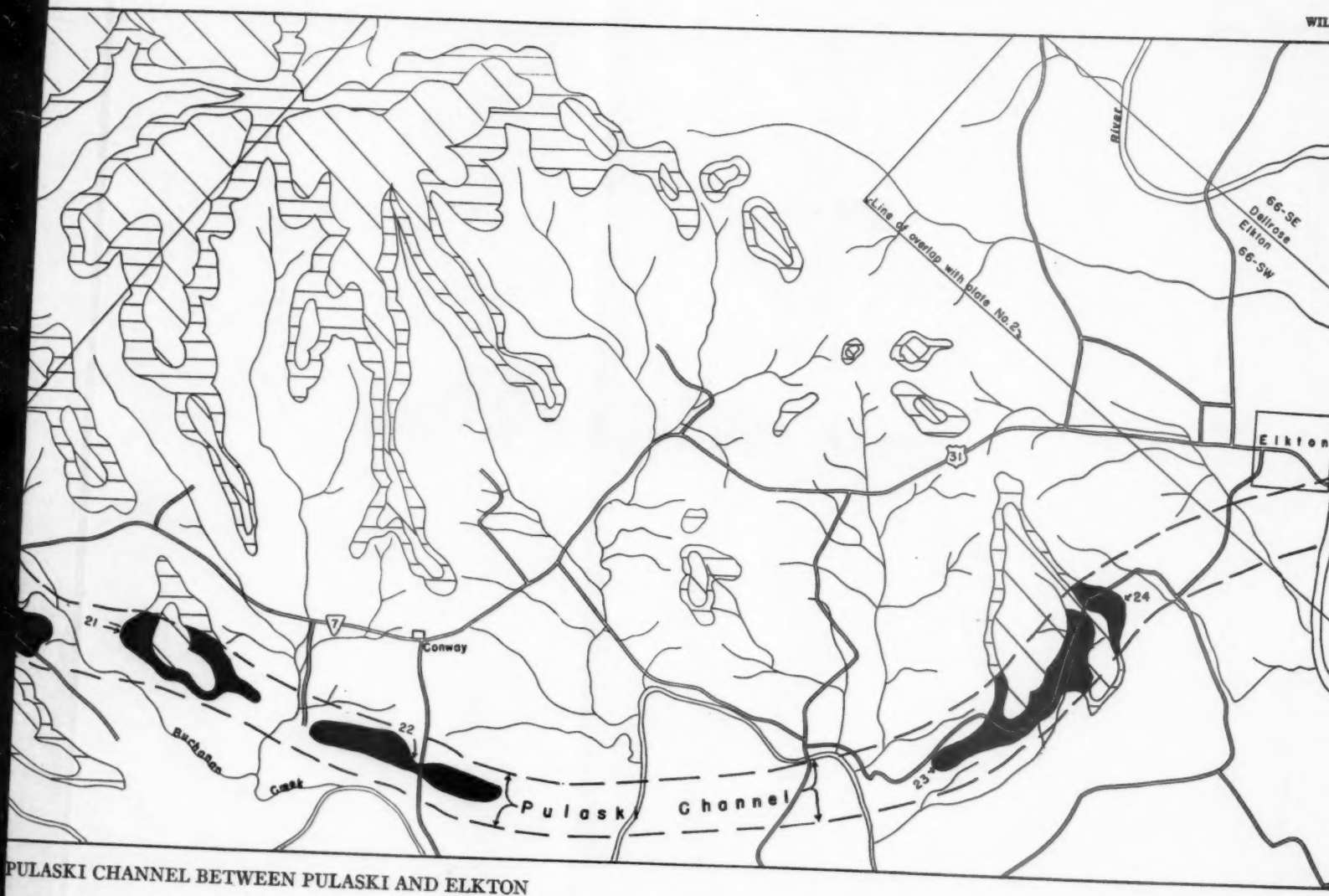



GEOLOGIC MAP OF THE VICINITY OF THE PULASKI CHANNEL BETWEEN PULASKI AND ELKTON




PULASKI CHANNEL BETWEEN PULASKI AND ELKTON

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 59, PP. 733-766, 2PLS., 11 FIGS. AUGUST 1948

# CHANNELS AND CHANNEL-FILLING SEDIMENTS OF RICHMOND AGE IN SOUTH-CENTRAL TENNESSEE

BY CHARLES W. WILSON, JR.

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## ABSTRACT

During the post-Maysville, pre-Richmond interval a stream flowing southeastward on the southeastern flank of the Nashville dome cut the Pulaski Channel in Giles and Lincoln counties, Tennessee, for a known length of 32 miles with average width of 1500 feet and depth of 100 feet. During Richmond time this channel was filled with clastic sediments brought in from the southeast by the advancing Richmond sea. About 75 feet of Lower shale and Conglomeratic sandstone was deposited during the first, or estuarine, phase. Lenses of Lower limestone conglomerate consisting of pebbles and cobbles of older underlying limestone in a matrix of sandstone were deposited during this phase. The second, or marine, phase began when the Richmond sea invaded the region and submerged the channel now three-quarters filled with estuarine sediments. Along the submerged channel a total of 70 feet of Upper limestone conglomerate, Upper shale, and Sandstone member of the Mannie shale was deposited contemporary with the accumulation of about 45 feet of normal Richmond over the surrounding region, consisting of Fernvale formation, Mannie shale, and Sandstone member of the Mannie shale. The Upper limestone conglomerate, which is equivalent to the Fernvale limestone member, was deposited upon the upper slopes of the channel and upon the top of the Lower shale. The submerged channel was filled flush with the surrounding sea floor by the Upper, or Mannie, shale.

At the same time the Fayetteville Channel in Lincoln County underwent a similar history, but the sediments filling this channel are too poorly preserved to permit satisfactory investigation and interpretation except by analogy with the Pulaski Channel.

## INTRODUCTION

This paper describes the thick clastic sediments of Richmond age that lie along an essentially continuous narrow belt in Giles and Lincoln counties, Tennessee (Fig. 1). The writer interprets this belt as an accumulation of sediments under, first, estuarine and, later, marine conditions in a deep stream channel cut during the post-Maysville, pre-Richmond interval. The sediments are believed to have accumulated during the early stages of the Richmond sea's invasion of south-central Tennessee and north-central Alabama when the Fernvale formation and the Mannie shale were deposited.

The first reference to these channels was made by Mehl and Pond (1929) who reported the "excavation of canyons to a depth of 150 feet or more about its (the Nashville dome's) southeastern margin in Lincoln County, Tennessee." Mehl

FIGURE 1.—*Index map of Giles and Lincoln counties, Tennessee, showing the location of the Pulaski and Fayetteville channels*

(1928) described the thick clastic sediments of Lincoln County and interpreted them as having been deposited in the channels or "Canyons." His description will be quoted verbatim later in this paper.

In 1940, while making a reconnaissance geologic map of the Central Basin of Tennessee for the State Division of Geology, the writer mapped the extent of the deposits in a general way. For several week-ends between 1940 and 1944, however, he went back to Giles and Lincoln counties for more detailed investigation.[1] During June 1944, while working for the Division of Geology, he completed the study of the deposits. This is essentially a reconnaissance study as no detailed mapping was done, but the writer believes that he can present a practically complete summary at this time.

## ACKNOWLEDGMENTS

The initial work for this project and some of the later work was done under the sponsorship of Walter F. Pond, former State Geologist of Tennessee. The writer expresses his appreciation to W. H. Bucher, H. B. Burwell, L. C. Glenn, W. B. Jewell, and R. R. Shrock who read the manuscript and offered useful suggestions. The writer also acknowledges the critical help of Ray Bentall, H. B. Burwell, Jack Collins, and W. B. Jewell during a 2-day field trip sponsored by the Tennessee Academy of Science in November 1945. The Natural Science Research Fund of Vanderbilt University paid for the final drafting of the figures and plates and for the typing of the text.

## PULASKI CHANNEL

### INTRODUCTION

Before proceeding with the detailed description of the belt of thick clastic sediments, the reasons will be set forth which support the conclusions that the deposits represent the filling of an old stream channel and not an infolded synclinal belt of younger strata. The origin and nature of the stream channel will also be discussed.

### CONTINUITY OF BELTS

The known occurrences of thick clastic Richmond in south-central Tennessee are aligned in a belt that extends from near Pulsaki, Giles County, southeastward to a point southeast of Coldwater, Lincoln County and is here designated as the Pulaski Channel. This narrow belt of sediments includes conglomerate, sandstone, shale, and thin layers of limestone conglomerate. These sediments indicate conditions of accumulation quite different from those under which the Richmond shale and limestone were deposited in this part of Tennessee.

The Pulaski Channel is first definitely recognized a short distance south of Pulaski[2]

[1] Part of the expense of this investigation was borne by the State Division of Geology, the rest by the writer.

[2] Two small and poorly preserved exposures of similar sediments occur 2 miles west of Wales and 1 mile southwest of Pulaski (Fig. 1). These are probably remnants of the Pulaski Channel.

and continues for 27 miles southeastward past Tarpley, Elkton, and Coldwater (Fig. 1). A composite of 16 miles, or about 60 per cent of the belt, consists of actual exposures of the unusual Richmond sediments; the remaining 11 miles consists of modern stream valleys cut below the Richmond channel-filling sediments, and of areas where the Richmond is still covered by younger sediments. The thick clastic Richmond in the vicinity of the Pulaski Channel consists of elongated exposures that line up with their long axes in the direction of the postulated belt. This fact together with the narrowness of the belt and its relatively great length proves that these thick clastic Richmond sediments occupy a channel that was formerly continuous. The types of sediments and their sequence are practically identical throughout the entire extent. Apart from four small scattered areas in Lincoln County discussed later in this paper, beds of similar nature aligned in this manner are not known elsewhere in south-central Tennessee.

### RELATION TO STRUCTURE

South-central Tennessee is located on the southeastern flank of the Nashville dome. This flank is modified by a super-imposed pattern of northwest-southeast structural "grain" consisting of an alternation of anticlines and synclines. These structures are known to have been active during the post-Catheys, pre-Maysville and the post-Maysville, pre-Richmond intervals by their control of the thickness and distribution of the Catheys, Leipers, and Richmond formations. Between Pulaski and Elkton the Pulaski Channel follows the axis of a broad asymmetrical syncline that has a relatively steep southwestern flank and a gentle dip on the northeastern flank. Between Pulaski and Elkton its position varies between 1 and 3 miles northeast of the steep southwestern flank. Between Elkton and the point southeast of Coldwater, where the belt disappears beneath younger sediments, however, it cuts across the anticlines and synclines paralleling the syncline which the upper part of the channel follows between Pulaski and Elkton. This proves that the agent controlling the origin of this belt was able to liberate itself locally from structural control and to cut across the structural grain.

### RELATION TO UNDERLYING STRATA

At practically all good exposures along the Pulaski Channel, it is clear that the thick clastic Richmond sediments were deposited within the confines of a narrow channel previously cut by some agent of erosion. The Leipers formation, which ranges between 25 and 75 feet thick along the Pulaski Channel, has been completely cut through, and the underlying Catheys formation, 50 to 125 feet thick along this belt, has been eroded to varying depths. Along U. S. Highway No. 31, about 1¼ miles south of Pulaski, erosion cut completely through the Catheys so that the channel-filling sediments locally rest upon the Bigby limestone which here underlies the Catheys.

At many localities the actual truncation of the essentially horizontal strata of Leipers and Catheys may be observed. The truncated strata are overlain by shale, sandstone, and conglomerate of the channel-filling sequence. A marked initial dip

often characterizes the basal portion of the channel-filling, indicating accumulation upon a relatively steep slope.

## ORIGIN AND NATURE OF CHANNEL

The sinuous, meandering areal pattern of the belt and the profiles of the vertical cross section across the belt and along the major dimension of the belt suggest to the writer that the channel was cut by a subaerial stream. No other agent of erosion seems to fit the facts as well. Several lines of evidence discussed later indicate that the stream flowed southeastward.

At the close of the Maysville age a widespread blanket of Leipers limestone covered Central Tennessee. That period of deposition was followed by differential uplift of the major positive structural units, or swells, of the region—that is, the Nashville dome and the southeastern extension of the Ozark dome (Wilson and Born, 1943). The differential uplift may have been positive vertical bulging of the two swells, passive stand of the swells with subsidence of the surrounding relatively negative region, or a combination of these two processes.

The differential uplift of the southeastern extension of the Ozark dome must have been appreciably greater than that of the Nashville dome as the upper part of the Trenton and all of the Maysville were stripped off this part of the State, resulting in Richmond strata overlying the Hermitage formation in Hardin, Wayne, and parts of neighboring counties. On the Nashville dome the Maysville was entirely removed only from Rutherford County and parts of adjacent counties. On the basis of regional structural control the region to the northwest of south-central Tennessee should have been relatively high; whereas, the region to the southeast should have been relatively low as it grades into the foreland slope of the Appalachian geosyncline. This relationship strengthens the belief that the stream flowed southeastward.

In addition to the large scale differential movement described above there was also small scale differential movement along the anticlines and synclines forming the structural grain on the southeastern flank of the Nashville dome. The relatively positive movement along the anticlines and the relatively negative movement along the synclines followed by erosion resulted in the preservation of an average of about 50 feet of Leipers in the synclines and only about 25 feet of Leipers on the anticlines and also in a slight topographic differential between the anticlines and synclines along any one cross section.

During the Maysville-Richmond interval, south-central Tennessee is visualized as a plain of very low relief and elevation, gently sloping south-southeastward, due to the greater relative uplift in the area which is now Hardin and Wayne counties. The continuity of this sloping plain was at first probably interrupted only by the very slight topographic differential of the structural grain. Although the elevation of the region is believed to have been low, it was sufficiently high for a stream to develop to the size required to cut a channel that had a known maximum depth of 100 feet and average width between 1000 and 2000 feet. The known length of the Pulaski Channel, the upper part of which followed the axis of a syncline that was slightly lower than the level of the adjacent and parallel anticlines, is 27 miles. The channel

probably continued at least 5 miles northwest of Pulaski to the northernmost known channel-filling sediments and undoubtedly extended many miles farther southeast beyond the point southeast of Coldwater where the channel-filling sediments today disappear beneath younger sediment. The gradient of the stream was undoubtedly very low. In cross section the ancient channel varies in the same way as channels of modern streams do, symmetrical throughout some cross sections, but with some slopes as steep as 20 degrees on the outside of meanders.

Since such a deep and wide channel was cut into an otherwise undissected plain, the cutting must have been relatively fast. If the cutting had been relatively slow, time would have been available for tributaries of comparable size with the master stream to have developed. Several near-by remnants of channel-filling sediments perched high above the general level of the channel are the only known evidence of tributaries. These seem to have been very short and had relatively steep gradients.

It is believed that the stream which cut this channel in a relatively short time would of necessity have carried a large volume of water. Its headwater area could conceivably have been many miles to the northwest, far up on the flank of the southeastern extension of the Ozark dome. The headwater tributaries of this stream might well have been instrumental in removing the Trenton and Maysville beds from that area.

On the bottom and slopes of the channel, which was cut through the Leipers formation into the Catheys limestone and locally even into the Bigby limestone, a rubble of granules, pebbles, and cobbles of these limestones accumulated. These were probably not carried downstream in great quantity but were rolled gradually down the gradient. Inasmuch as the rocks over which the stream flowed were dominantly limestones with no material coarser than silt and clay, only minor quantities of this grade of clastic material were carried by the stream. Most of the load of sediments carried was probably in chemical solution as calcium bicarbonate.

### ORDOVICIAN SYSTEM

*Trenton group.*—The oldest beds reached by the downcutting stream are in the *Bigby limestone.* This limestone is blue-gray when fresh and brown-gray when weathered. Leaching leaves a porous, brown, phosphatic residuum. The formation is massive-bedded, and the horizontal bedding planes are difficult to recognize because of the predominance of cross-bedding planes. It is coarse-granular in texture, composed of ground fragments of fossils and of crystalline grains of calcite. Near the channel the formation ranges between 60 and 80 feet thick. Fossils are relatively rare.

In only one area—1¼ miles south of Pulaski—the Bigby limestone was reached by the downcutting stream.

Throughout most of the Pulaski Channel, the lower part of the channel-filling sediments rests upon beds of the Catheys formation. This formation contains a variety of lithologic types but locally consists chiefly of blue-gray limestone in beds ranging between 2 and 8 inches thick. Typical surfaces of these beds are irregularly knobby giving the limestone a general nodular appearance. Thin partings of gray shale

separate the beds of limestone. Near the channel, the Catheys formation ranges between 50 and 125 feet thick. Common fossils are *Columnaria alveolata*, *Tetradium fibratum*, *Stromatocerium pustulosum*, *Constellaria teres*, *C. emaciata*, *Heterotrypa parvulipora*, and *Platystrophia precursor* and varieties.

*Maysville group.*—The upper slopes of the channel were cut into beds of the *Leipers formation*. Lithologically, this formation resembles the underlying Catheys very closely, but the fauna is distinctly different. In the vicinity of the Pulaski Channel the formation ranges between 25 and 75 feet thick. Among the numerous fossils that characterize the Leipers are *Hallopora rugosa*, *H. ramosa*, *Heterotrypa frondosa*, *Platystrophia sublaticosta*, *P. laticosta*, and *P. ponderosa*.

*Richmond group, normal regional sequence.*—Post-Richmond erosion removed all

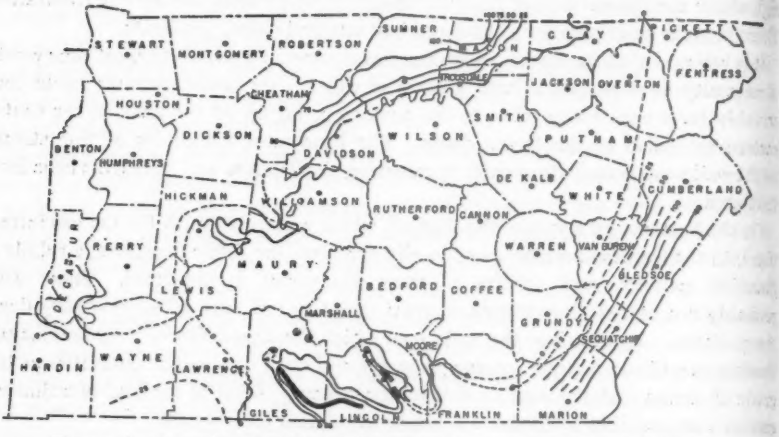


FIGURE 2.—*Isopach map of Central Tennessee, showing the variation in thickness of the Richmond group*

This map was compiled from data collected for the Tennessee Division of Geology. The contour interval is 25 feet.

sediments of Richmond age from south-central Tennessee, except where locally protected by relatively low structural position. The preserved remnants occupy two broad synclinal areas in Giles and Lincoln counties, the southern margin of Lincoln County, and several small isolated remnants in the northern parts of these counties (Fig. 2).

The normal regional sequence of Richmond units, exclusive of all phases of the thick clastic sequence limited to the channel, is described in the following paragraphs. This entire sequence accumulated under normal shallow-water marine conditions.

FERNVALE FORMATION: The lowest unit of the Richmond in south-central Tennessee is the *Calcareous mudstone member*,[3] a zone of grayish-green mudstone and siltstone that is ordinarily less than 10 feet thick, characterized by the persistent occurrence of *Dalmanella meeki*. It is exposed in east-central Lincoln County and in central

[3] This informal descriptive name was proposed by the writer (1948).

Franklin County. In the Sequatchie Valley, Marion County, it grades into the lower part of the thick Sequatchie formation. The unit represents the sediments deposited during the early invasion of south-central Tennessee by the Richmond sea and that were subsequently overlapped by the much more extensive Fernvale limestone member. The Richmond seas invaded Central Tennessee on both the southeastern and northwestern flanks of the Nashville dome simultaneously as the same unit is well developed in the northwestern half of Davidson County and western Sumner County.

The *Fernvale limestone member* (Fig. 2) consists of massive and irregularly bedded, coarsely crystalline limestone in vari-colored combinations of red, pink, green, and gray. *Rhynchotrema capax* is one of the most common fossils. In south-central Tennessee it ranges from less than 10 up to 30 feet thick. During its deposition, the Richmond sea reached its maximum extent in Central Tennessee, covering all of the region.

MANNIE SHALE: The Mannie shale is ordinarily green although at some localities it is bluish gray. It is locally a clay-shale that breaks into small, thin flakes. Fossils are relatively rare. In south-central Tennessee it is as thick as 25 feet. Probably the best exposure of this unit is on the east slope of Ingram Hill overlooking Buchanan Creek on U. S. Highway No. 64 about 5 miles east of Pulaski.

This formation is apparently conformable with the underlying Fernvale limestone member. Its areal extent is less than that of the Fernvale limestone, as it has been locally removed by post-Richmond erosion.

The existence of the *Sandstone member of the Mannie shale* was first recognized by Born and Wilson (1939) and referred to as "local sandstone lenses." The present writer has found this unit to be more extensive and to be the uppermost unit of the normal regional Richmond sequence, occurring at several localities many miles from the Pulaski Channel. In present stratigraphic nomenclature it is considered a member of the Mannie shale with which it is closely associated.

This member is a very fine-grained, uniformly textured quartz sandstone ordinarily about 10 feet thick. It is white when fresh and buff or brown when weathered. Its limited area is the result of removal by post-Richmond erosion.

Most of its exposures are in the vicinity of the Pulaski Channel where they were preserved from post-Richmond erosion by local lower structural position due in part to compaction of the underlying thick shale, but these exposures are not restricted to the limits of the channel as the Sandstone member at some localities can be traced across the top of the thick clastic, channel-filling sequence to its stratigraphic position in the normal regional Richmond sequence. This is well demonstrated between Localities 12 and 13 (Pl. 1) and Figure 5. Two good exposures of this unit at some distance from the channel are (1) only a short distance southeast of Riversburg, about 4 miles north of Pulaski, and (2) about 3 miles east of Lynnville or 2 miles east of Happy Hill in northeastern Giles County.

Marine fossils were found near the base of the Sandstone member at Locality 26 (Pl. 2), about 1 mile southeast of Elkton.

The Sandstone member of the Mannie is believed to be conformable with the Mannie shale and to represent the last beds deposited in Central Tennessee during the

inundation by the Richmond sea. It is overlain unconformably by the Brassfield limestone of Silurian age or by Mississippian strata where the Brassfield has been removed by erosion.

*Richmond group; thick, clastic, channel-filling sequence.*—One purpose of this paper is to show that the normal regional Richmond and the thick clastic channel-filling Richmond are contemporary deposits. As they are very different in significant details, it is possible to correlate the individual units in only a general way. One difficulty is the presence in the channel-filling Richmond of several units which are absent in the regional Richmond. Insofar as possible, however, the names of the units of the normal regional sequence will be used for the channel-filling sediments.

Before giving the detailed evidence on which the writer's view is based he wishes to quote Mehl's (1928) reasons for thinking that the channel was cut through normal regional Richmond and that the channel-filling sediments are therefore younger than the regional sediments—Silurian rather than Ordovician (*i.e.*, Richmond). Mehl proposed the name Camargo formation for the thick clastic sediments and described it as follows:

"This formation consists predominantly of greenish blue to olive or light-brown arenaceous shales, and shaly sandstones. In several places, however, the basal members consist of a considerable thickness of limestone conglomerate and heavy-bedded, coarse, brown sandstone. In both lithology and thickness this formation is markedly variable in short distances; but because of its peculiar relation, it is readily distinguishable from other formations for the most part.

"The lower members of the Camargo are distinctly valley fills, a fact that accounts for the variations in lithology, and in part, the variable thickness. The coarse limestone blocks of the talus along the valleys of the Pre-Camargo surface now constitute the Camargo conglomerates that are 'plastered' across the eroded edges of Fernvale, Leipers, and Catheys. This conglomerate grades laterally toward the center of the old valleys into coarse sandstone and even shales. Succeeding the 'fills' proper is a considerable thickness of light-brown to olive or green, arenaceous shale that is much more widespread. In some places this part of the formation rests directly on the Fernvale or older rocks; but, in regions where the lower fill is present, there is no evident break between the two parts of the formation.

"Good exposures of the Camargo conglomeratic phase are to be found in several places in Good Hollow and branches, about two miles southwest of Camargo, and again, about one-half mile west of Coldwater on the Coldwater-Blanche road. Other outcrops that emphasize the irregularity of distribution and the possibilities of many buried channels beneath the Highland Plateau area of the southern part of the county are those about two miles northeast of Blanche and 1½ miles north of Cash Point.

"The thicknesses of the fill appear to vary between 30 feet and something over 100 feet and are measured from the lowest exposure to the highest adjacent Fernvale outcrop. The upper, more strictly shaly part of the formation, is well exposed at the heads of steep valleys about one-half mile north of Howell. Here the thickness of the shale between the Fernvale and the Chattanooga formation, approaches 200 feet. It seems likely that total thicknesses of as much as 350 feet may be expected in wells that encounter some of the buried channels.

"Limited search has failed to produce fossils from this formation other than those contained in the blocks of the limestone conglomerate and some phases of the limestone matrix of the conglomerate. In some regions blocks of Catheys, Leipers, and Fernvale are imbedded in a mass of coarse limestone debris that contains Fernvale fossils. It is evident, however, that these shells are part of residual material from the weathering of Fernvale and not the remains of organisms living in the Camargo sea invasion. None of these shells is found in the lateral gradation of the conglomerates into sandstone or shale, transitions that take place within a horizontal distance of as little as 30 feet from the walls of the old valleys."

Mehl summarized the history of the channels and channel-filling thus:

"The interval during which the Fernvale limestone was deposited was brought to a close by the elevation of the Nashville dome considerably above the sea. Toward its southeastern margin, the Lincoln Cou..y area, its elevation was sufficiently great to permit the formation of deep valleys. At the time of this elevation, the sediments of the Fernvale formation were but little consolidated

and the streams cut rapidly through to the more resistant Leipers and Catheys. As is evidenced in Lincoln County, there was soon developed a system of steepsided valleys some of which were considerably more than 100 feet deep. There accumulated over these slopes considerable masses of talus which included large and small blocks of Catheys and Leipers limestone and more or less unconsolidated masses of Fernvale. The surface of the high flats and ridges was probably covered with a thick mantle of sand, clay, and limestone fragments from the recently deposited Fernvale. This was the nature of the surface at the time of the next sea invasion in Camargo times.

"Deposition in the Camargo sea

"During Mid-Silurian times or perhaps slightly later, the sea again invaded this area. During its landward progression the waves reworked the talus covering of the valley walls and swept out, to settle in the deeper parts of the channels, much of the materials of the coarseness of sand and finer. As a consequence, the coarse material along the sides of the valleys graded laterally into sands and clays. The fossils that had weathered from the Fernvale limestones and shales were too large to be transported far and they with coarse limestone waste settled into the interstices between the large limestone blocks of the talus deposits.

"After the channels had been filled, the sea extended over even the higher parts of the region; but the water was so shallow on land so near to the north that clays and fine sand were distributed uniformly over valley fills and Fernvale knobs. In some places there was built up a thickness of more than 200 feet of sandy shales over the divides between the valleys of the Post-Fernvale stream system.

"Apparently there was little or no life in this advancing sea for no fossils are to be found in the sands and clays that filled the deeper parts of the valleys. Even after the channels had been filled and the sea extended over the higher points of land, the rapidly accumulating mud and sand prevented the sea life from spreading to this region."

Calcareous Mudstone Member Of The Fernvale Formation: This unit is not present in the Pulaski Channel at least as far southeast as a point 2 miles southeast of Coldwater. It is believed, however, that farther down the channel beneath the cover of Mississippian strata this unit is present at the base of the channel-filling sequence, deposited there as an estuarine deposit within the limits of the channel when the early Richmond invasion encroached upon Central Tennessee from the southeast, reaching much farther inland in the Pulaski Channel as an estuary. It is not known how far such estuarine conditions existed up the channel, but certainly not as far as the point 2 miles southeast of Coldwater (Pl. 2, loc. 39), where there is no suggestion of the member below the definite thick clastic channel-fill.

Conglomeratic Sandstone: This is the lowermost unit of the thick clastic sequence known in the Pulaski Channel. Stratigraphically it is probably part of the early Fernvale formation not found outside of the channel in Tennessee, but more widely represented in Alabama. In cross section it consists of a meniscoid, or concavo-convex, lens (Fig. 3) that average about 1000 feet wide. It extends throughout the length of the Pulaski Channel. The maximum recorded thickness in the center of the lens is 35 feet. Along the slopes of the channel it often has initial dips up to 20°. At some localities the edges of the lens rise as much as 20 feet above the top of the main body of the unit.

The Conglomeratic sandstone consists of coarse ($\frac{1}{2}$ to 1 mm) and very coarse (1 to 2 mm) quartz sand that contains granules (2 to 4 mm) of quartz. Occasional grains are larger than 4 mm in diameter. Dr. W. B. Jewell examined the sand and reported that it consists of grains of unstrained quartz that are unusually well rounded and frosted.

The unit is very massive-bedded, with cross-bedding dominating, some units of the cross-bedding being as thick as 3 feet, and, as far as observed, most of the foresets dip north-northwestward between Pulaski and Elkton and westward at all exposures in the Pulaski Channel east of Elkton. These directions are upstream with reference

FIGURE 3.—*Diagrammatic vertical cross section across the channel-filling sediments to show the stratigraphic units and their relationships*

to the direction the eroding stream flowed; landward with reference to the invading sea.

The interstices between the grains and granules of quartz are filled with crystalline calcite that forms the matrix for the Conglomeratic sandstone. The matrix contains sufficient ferruginous and other foreign material to color the unit dark brown after weathering. The calcareous material is often gray and green when fresh. As it leaches readily it has been removed from all the sandstone that is not freshly exposed.

The areal pattern and vertical cross section of the Conglomeratic sandstone suggest either fluviatile deposition of clastic sediments from the north or northwest or accumulation under estuarine conditions of clastic sediments washed up the estuary by tidal currents at a time early in the Richmond invasion when only the lower part of the channel was submerged. The latter hypothesis is believed to fit the following observed facts and interpretations.

(1) The matrix of limestone suggests marine or brackish-water environment rather than fresh water.

(2) The cross-bedding faces up-channel suggesting transportation of the sand in general from southeast to northwest, opposite to the postulated southeastern drainage of the stream that cut the channel.

(3) There is an available source for these sediments to the southeast as grains of quartz similar in appearance, size, and occurrence (*i.e.*, imbedded in a limestone matrix) to those found in the Pulaski Channel occur to the southeast near Collinsville, Alabama, in Fernvale limestone (Butts 1926, p. 129, Pl. 29 B). Grains and granules of quartz in samples collected there by the writer were identical, as far as observed, with those found in the Pulaski Channel. On the other hand, there is no known source for such clastics from the northwest.

Strong up-channel tides and currents are believed to have carried the sand northwestward up the estuary as far as what is now the northern line of Giles County in spite of receding tides and of fresh water flowing down the drowned stream channel. The predominant advancing tidal currents pushed the sand northwestward with the foreset beds of cross-bedding facing northwestward in general. Receding tides and the fresh-water current were unable to stop the advancing sand, but did scour off and rework the sand forming the topset beds and the upper part of the foreset beds, producing typical truncated cross-bedding.

LOWER LIMESTONE CONGLOMERATE: This unit is very closely associated with the Conglomeratic sandstone, locally occurring below it, as between Localities 2 and 3, at Locality 7, (Pl. 1), and at Localities 25, 35, and 39 (Pl. 2) and sometimes interbedded with it (Pl. 2, loc. 42).

The term "Lower" distinguishes this unit from a somewhat similar limestone conglomerate that was formed and is associated with younger units than the Conglomeratic sandstone. The younger, or "Upper", limestone conglomerate will be discussed later in its stratigraphic position.

The Lower limestone conglomerate consists of subangular and rounded pebbles, cobbles, and locally even boulders of Bigby, Catheys, and Leipers limestone in a matrix of Conglomeratic sandstone with its characteristic coarse grains and granules of quartz in a matrix of crystalline limestone. It occurs as local lenses beneath the

Conglomeratic sandstone in the bottom of the channel, interbedded with its basal zones, or plastered on the truncated edges of the older underlying limestone and beneath the Conglomeratic sandstone (or the lower part of the Lower shale as will be explained later) on the slopes of the channel. The latter occurrences are restricted to the lower slopes of the channel, not many feet above the top of the thinned edge of the Conglomeratic sandstone.

In one of the most unusual occurrences (Pl. 2, loc. 42), three beds of Lower limestone conglomerate, each averaging about 4 feet thick, are interbedded with Conglomeratic sandstone beds of similar thickness within the basal 25 feet of the Conglomeratic sandstone.

Apparently, prior to the beginning of flooding of the stream valley by brackish water the bottom and slopes of the Pulaski Channel had an irregular accumulation of pebbles and cobbles of older limestone either as pockets or lenses on the bottom or as patches perched upon the slopes. The estuarine invasion from the southeast that brought in the grains and granules of quartz which formed the Conglomeratic sandstone and that flooded only about the lower third of the vertical cross section of the Pulaski Channel reworked the pebbles and cobbles and redeposited them within the lower third of the channel as lenticular-shaped bodies. Those pebbles and cobbles of limestone perched above the lower third of the channel were not affected at this time except, of course, individual pebbles and cobbles that may have rolled down the slopes of the channel into the water in the estuary under the influence of gravity and slope-wash.

Shale Unit: Overlying the Conglomeratic sandstone is a thick sequence of about 100 feet of shale that cannot everywhere be subdivided. At several localities, including Localities 3, 6, 12, 16, 18, 29, 33, and 37, beds of Upper limestone conglomerate are intercalated *into* the approximate middle of this shale sequence. Near the channel wall these beds of Upper limestone conglomerate average about 3 feet thick and contain pebbles and cobbles of older underlying limestone and some grains of quartz. They dip downward toward the center of the channel, due to initial dip and subsequent compaction, and grade into very thin beds of Upper limestone conglomerate that contain very small pebbles and granules of older underlying limestone and more numerous quartz grains. Where beds of Upper limestone conglomerate do not occur the entire thickness of shale is unbroken. Where they do occur they divide the shale into an underlying, or "Lower shale", and an overlying, or "Upper shale". As will be explained in more detail later the Lower shale is believed to be estuarine in origin, having accumulated in the channel before the Richmond sea flooded the region, and the Upper shale is believed to be marine in origin, being correlated with the Mannie shale of the normal regional sequence.

*The Lower shale* overlies the Conglomeratic sandstone with apparent conformity and is similarly restricted to the lower part, or bottom, of the channel. It extends slightly beyond the Conglomeratic sandstone, to rest direcly upon the truncated older underlying limestone. The shale has sufficient initial dip near the edge of the channel to rise several feet up the slope of the channel. In vertical profile it is a concavo-convex lens. Its thickness, where a bed of Upper limestone conglomerate is present so that its thickness can be measured, ranges between 20 and 50 feet.

At some localities lenses of Lower limestone conglomerate occur between the Lower shale and the older underlying limestone. These are interpreted as lenses formed by the reworking and redeposition of the pebbles and cobbles of older underlying limestone and grains of quartz on the slopes of the channel by the progressively rising water level in the estuary. This conglomerate has the same matrix of sandstone as the Lower limestone conglomerate associated with the Conglomeratic sandstone. The source of this matrix would have been from the sand washed by high tides or blown by the wind higher up the slopes of the channel than the top surface of the Conglomeratic sandstone. At some localities today the meniscus of the Conglomeratic sandstone rises over 20 feet above the main level of the sandstone. The exposure between Localities 36 and 37 (Pl. 2) illustrates occurrences of Lower limestone conglomerate associated with the Lower shale.

The shale is fairly well laminated and is gray to green. It contains lenses of fine-grained sandstone as thick as 3 feet (Pl. 1, loc. 6; Pl. 2, loc. 29 and 33) ranging to small ones about 1 inch thick and 6 inches or so in diameter. Some of the latter show a spiral pattern that suggests the swirling action of eddies in the concentration of sand grains on the muddy bottom of the estuary. The fine sand of the Lower shale is also concentrated in thin laminations that are not obvious when fresh. Upon weathering, however, they turn brown and are cemented into hard, thin flakes that accumulate on the surface of the glades. Fucoidal markings are often present on the underside of these flakes, standing in bas-relief above the surface suggesting mud castings of some bottom-living animal or the filling of burrows made by these animals. Another interesting occurrence of sand is in elongated plano-convex lenses that suggest the filling of scour channels or rills. These range from about 1 inch up to 6 inches wide. Occasional grains and granules of quartz and granules and small pebbles of older limestone occur in beds of the Lower shale unit.

At the close of deposition of the Lower shale unit the estuary had received about 35 feet of Conglomeratic sandstone and an average of 40 feet of Lower shale, ranging between 20 and 50 feet thick depending upon the location in the channel. These sediments were restricted to the lower part, or bottom, of the channel as so far the Richmond sea had not spread beyond the limits of the Pulaski Channel. This channel, originally about 100 feet deep, was still about 25 feet below the level of the surrounding region, and as yet estuarine conditions existed in the channel and marine deposition was restricted to the region southeast of the preserved part of the sediments that filled the Pulaski Channel.

After the estuarine accumulation of the Lower shale, beds of the *Upper limestone conglomerate* were locally deposited upon the Lower shale or upon the older limestone forming the upper slopes of the channel. These beds are believed to indicate that after 20 to 50 feet of silt (Lower shale) had been deposited in the channel, regional invasion started, first flooding the channel and then spreading over all of south-central Tennessee. The gradual encroachment of the Richmond sea must have been accelerated so that the remaining 25 feet of the channel was quickly flooded and submerged. The widespread regional distribution of the relatively thin Fernvale limestone indicates a relatively rapid invasion by the sea. All subsequent Richmond deposits in south-central Tennessee were marine rather than estuarine.

The abrupt invasion produced different conditions of deposition from those under which the Conglomeratic sandstone and Lower shale had been deposited. The limestone floor of the region was now flooded by a shallow, clear water sea, whereas the submerged channel was a submarine belt covered by this sea, different only in being about 25 feet deeper and in having a muddy bottom of unconsolidated silt. No break in deposition is visualized at the close of the accumulation of the Lower shale, but rather a relatively abrupt change in the conditions of sedimentation. Among these changes in the sedimentation within the channel were:

(1) Change from brackish-water estuarine conditions to marine conditions.

(2) Practically no silt was now being carried from the southeast to the region which is now Tennessee, so that the water in the channel was practically free of all silt except that which was being reworked from the silt previously deposited in the channel (the Lower shale).

(3) Reduction of tidal action, although the tides in the submerged channel were still stronger than those in the shallow Fernvale sea.

(4) All of the remaining debris of older underlying limestone on the slopes of the channel was soon incorporated into beds and lenses of limestone conglomerate (Upper) contemporary with the Fernvale limestone.

During the time that the clear water sea, in which the Fernvale limestone was deposited, existed throughout the region, sedimentation occurred in the now-submerged channel, but it is impossible to delimit and to correlate the sediments deposited in the channel with the Fernvale limestone with any close degree of detail. In general, it is postulated that all the beds of Upper limestone conglomerate and whatever shale is interbedded with several beds of Upper limestone conglomerate, as at Locality 16 (Pl. 1), are correlative with the Fernvale limestone of the normal regional sequence.

The better-exposed occurrences of the Upper limestone conglomerate are as follows:

(1) Occurrences between the Lower shale and Upper shale near the middle of the channel:
Localities 3 and 6 (Pl. 1), about 1¼ miles south of Pulaski
Locality 16 (Pl. 1), about 1½ miles southeast of Tarpley
Locality 33 (Pl. 2), about ½ mile west of Coldwater
(2) Occurrences plastered high on the sides of the channel:
Localities 29 and 30 (Pl. 2), in Reed Branch
Locality 32 (Pl. 2), about 1 mile southwest of Coldwater
Locality 41 (Pl. 2), less than 1 mile north of the eastern end of the Pulaski Channel
(3) Occurrences at the edge or shoulder of the channel:
Locality 4 (Pl. 1), about 1½ miles south of Pulaski
Locality 37 (Pl. 2), about 1½ miles southeast of Coldwater

Ordinarily, at the edge, or shoulder, of the channel the Upper limestone conglomerate is a bed 2 or 3 feet thick consisting of pebbles and cobbles of the older underlying limestone and quartz grains in a matrix of Fernvale limestone. Here it usually rests upon the Leipers formation. Toward the middle of the channel the limestone beds have often disappeared, but where present they are only a few inches up to a foot thick and contain only a few rounded granules of older limestone and varying amounts of quartz grains. They have a well-preserved marine fauna and are similar to normal

Fernvale limestone paleontologically and lithologically except that they contain quartz grains and limestone granules.

Unfortunately, exposures do not permit tracing the Upper limestone conglomerate from the edge of the channel into normal Fernvale limestone of the surrounding region. If such exposures had been preserved it would be easy to prove or disprove the interpretation, here proposed, that the Upper limestone conglomerate at the edge of the channel would grade into normal Fernvale limestone outside the channel and slightly above the level of the edge of the channel.

It is believed that contemporaneous with the deposition of the Upper limestone conglomerate, and any shale that might be interbedded with it, as at Locality 16, Plate 1, Fernvale ooze was being widely deposited over south-central Tennsessee. On this basis all beds below the Upper limestone conglomerate would be early Fernvale beds, widespread to the southeast in Alabama but restricted in Tennessee to the estuarine deposits of the Pulaski Channel. The Calcareous mudstone member of the Fernvale formation is the only earlier Richmond unit—that is, pre-Fernvale limestone —present in the normal regional Richmond sequence in south-central Tennessee. The Conglomeratic sandstone and the Lower shale either (1) were deposited during a brief pause of deposition or (2) they represent a wedge of very rapidly accumulated sediments deposited contemporaneously with continuous deposition elsewhere during the change from the slow accumulation of silt of the Clacareous mudstone member to the still slower accumulation of the ooze that formed the Fernvale limestone member.

The ooze that later formed the normal Fernvale limestone extended to the edge of the channel at the level of the region. The Fernvale limestone ordinarily ranges from 2 to 10 feet thick in the immediate vicinity of the channel. Just at the break of the channel, slightly below the level of the region and still overlying the older underlying limestone, the average thickness of ooze was less than that on the regional level due to the exposure of this position to submarine scour. In addition to being thinner, the ooze had incorporated in it the pebbles and cobbles of older underlying limestone and any quartz grains present that high on the slopes of the channel. This coarse conglomeratic phase is usually in the basal part and is overlain by more or less normal Fernvale limestone with a reduced amount of conglomeratic material. The zone of ooze with its imbedded pebbles and cobbles extended obliquely out and down into the channel over the surface of the Lower shale. The farther toward the middle of the channel it extended the thinner it became and the smaller the included particles. In some places, the Upper limestone conglomerate is only 1 or 2 inches thick and at other places it was not deposited at all.

About $1\frac{1}{4}$ miles southeast of Tarpley (Pl. 1, loc. 16), two or possibly three thin beds of Upper limestone conglomerate occur above at least 40 feet of Lower shale in a position near the middle of the channel. The lower one is about a foot thick and is separated from the one next above by 15 feet of shale. The limestone conglomerate beds are lenticular and consist of a matrix of Fernvale limestone, lithologically and paleontologically similar to normal Fernvale, containing a large amount of sand and some granules of older underlying limestone. The 15 feet of shale between the two beds of limestone is considered as contemporary with the Fernvale limestone of the region, the source being silt reworked from older parts of the channel-filling and re-

deposited by the Fernvale sea in the lowest part of the channel. About 15 feet above these beds, and occupying the top of the hill (Pl. 1, loc. 16), a floor of slabby calcareous sandstone probably represents a third zone of Upper limestone conglomerate. As some doubt can be raised as to whether it is in place, it will not be considered beyond this brief mention.

Half a mile west of Coldwater (Pl. 2, loc. 33), about 7 feet of Upper limestone conglomerate occurs approximately 25 feet above the top of the Conglomeratic sandstone. Some beds of this zone are typical Fernvale limestone except that they include many sand grains. Other beds contain rounded pebbles and cobbles of older underlying limestone. At this place the deposition occurred near the middle of the channel, and the pebbles and cobbles were rolled from the wall of the channel to the point of deposition by the Fernvale sea.

In the west fork of Reed Branch (Pl. 2, loc. 29), a lens of about 3 feet of coarse-textured Upper limestone conglomerate wedges out with surprising abruptness, suggesting an actual rock slide from the channel wall out several feet into the channel upon the top of the Lower shale. The rubble of limestone debris on the steep upper slope of the channel was submerged for the first time, and ooze of Fernvale limestone began forming in the interstices and over this rubble. Gravity and tides loosened the water-saturated mass, causing it to flow glacierlike to a lower level of repose upon the Lower shale.

The contrast of contemporary sedimentary environments is difficult to visualize, but it seems to the writer that this is the only possible explanation of the correlation of the beds of the Upper limestone conglomerate near the middle of the Shale unit and the Upper limestone conglomerate on the edge of the channel with the normal regional Fernvale limestone. The concentration and accumulation of thin beds of limestone conglomerate (Upper) and some silt in the channel while widespread Fernvale limestone was being deposited over the surrounding region is believed to be the result of the following factors.

(1) During the accumulation of the sand of the Conglomeratic sandstone and the silt of the Lower shale, some of the sand and silt was left as veneer on the sides of the channel above the mean level of the main body of Conglomeratic sandstone and Lower shale. This would have resulted from the meniscus of initial dip, the wash of exceptionally high tides, and wind work during periods when exceptionally low tides permitted the local exposure and drying of sand and silt. Post-Lower shale and pre-Fernvale limestone compaction would have supplemented the differential in position between the sand and silt on the upper slopes of the channel and the top of the Lower shale in the middle of the channel. All the sand and silt on the upper slopes of the channel would have been redeposited in the channel by the rapidly rising Fernvale sea, thus supplying some silt for deposition contemporary with the accumulation of normal Fernvale ooze over the region.

(2) Waves, tides, and currents of the Fernvale sea would have scoured the upper part of the silt of the Lower shale, and the reworking and redeposition of this silt in the channel would have continued during the early part of the Fernvale invasion. As the tidal currents within the limits and immediate vicinity of the submerged

channel were relatively stronger than those in the surrounding sea, any silt present in its immediate vicinity would have been concentrated along the channel.

(3) The silt-covered bottom of the submerged channel was an unfavorable environment for marine life of the "shelly facies" type and thus inhibited the general accumulation of calcareous ooze within the limits of the channel.

Accumulation of the calcareous ooze, which became the matrix of the Upper limestone conglomerate, on the relatively steep slope of the estuary and upon such an unsteady surface of deposition as a rubble of pebbles and cobbles of older limestone would be accompanied by a marked amount of contemporary deformation. Even sticky chunks of Fernvale ooze, itself, may have been freed to slide down the slope to occur today as partially rounded cobbles of Fernvale limestone imbedded within an essentially contemporary matrix of the same material. Possible examples of such contemporary deformation of the Fernvale limestone and Upper limestone conglomerate occur at Localities 29, 30, 32, 33, and 41 (Pl. 2). Such occurrences would represent slides that moved a few feet down the slope and even out into the channel for several feet upon the surface of the Lower shale. The abrupt ending of the wedge of Upper limestone conglomerate at Locality 29 is believed to have resulted from such a local slip.

Following the accumulation of the Fernvale limestone over the region and the Upper limestone conglomerate within the submerged channel, the clear water conditions of the sea were interrupted by the addition of large quantities of silt. This silt—the Mannie shale of the normal regional Richmond sequence—was spread over the region to form a blanket with an average thickness of 25 feet. In the submerged channel, however, which was still nearly 25 feet below the regional level, this silt accumulated to an average thickness of 60 feet in the deepest part of the channel, with a range between 50 and 70 feet. At the close of this phase of deposition the channel was filled level with the silt deposited over the surrounding region. This shale unit can be separated from the Lower shale only if a bed of Upper limestone conglomerate is present.

This *Upper shale* is lithologically similar to the Mannie shale of the region. It is a thinly laminated, greenish clay-shale. In the channel it contains occasional concentrations of fine sand in lenses or thin beds and scattered coarse grains and granules, but these are relatively rare in comparison to the concentration in the Lower shale.

The closing phase of the Richmond age was the widespread deposition of about 10 feet of very fine-grained sand—the *Sandstone member of the Mannie*, or *Upper shale*—uniformly and conformably over the relatively thin Mannie shale of most of south-central Tennessee and the relatively thick Mannie shale of the filled channel. The occurrences of this unit within the limits of the channel are similar to those outside of the channel in lithologic character, thickness, and stratigraphic position. After the sand was deposited the Richmond sea withdrew from south-central Tennessee.

## SILURIAN SYSTEM

*Medinan series.*—During the Richmond-Brassfield interval the area was subjected to subaerial erosion that removed much of the Sandstone member of the Mannie shale

and some of the Mannie shale itself. The preservation of about 90 per cent of all the Sandstone member within the limits of the filled channel is probably the result of favorable, relatively low, structural position produced locally by differential compaction of 100 feet of the Shale unit below the Sandstone member.

As a result of the invasion by the Brassfield sea, the *Brassfield limestone* was deposited as a blanket over south-central Tennessee, resting unconformably upon the Sandstone member of the Mannie shale where it is present, or upon the underlying Mannie shale where the Sandstone member had been removed. Locally, even the Mannie shale had been eroded, and the Brassfield rests on the Fernvale limestone member in the regional stratigraphic succession, but not within the limits of the channel. The Brassfield overlies the Richmond throughout practically the entire length of the Pulaski Channel with the major exception of the southeasternmost 3 miles where the Brassfield is locally absent and the Richmond is overlain by the Chattanooga shale.

The Brassfield consists of thinly bedded, blue-gray, finely crystalline limestone containing many species of corals. Dense black to brown chert is present in exposures of fresh limestone, and during weathering the amount of chert apparently increases until, after all the limestone has been leached, a sequence of thin layers of chert is left. Along the Pulaski Channel the thickness varies up to a maximum of 45 feet.

As is so well illustrated between Localities 12 and 13 (Pl. 1; Fig. 4), between Localities 29 and 31 (Pl. 2; Fig. 6), and at Localities 21 and 24 (Pl. 1), and 26, 32, and 33 (Pl. 2), along the Pulaski Channel normal Brassfield limestone extends completely across the channel, indicating that the final chapter of the channel-filling had been finished long before the invasion of the Brassfield sea.

*Niagaran series.*—At two localities along the Pulaski Channel there are small remnants of these Niagaran formations which have been locally preserved from erosion. These two outliers are remnants of the more extensive blanket of Osgood, Laurel, and probably younger Niagaran formations that covered this area before being removed by pre-Chattanooga erosion. One of the localities is the hill between Buchanan Creek and U. S. Highway No. 31 about $1\frac{1}{4}$ miles northwest of Conway. The other is in the headwater region of Kelly Creek about 2 miles west of Blanche.

The *Osgood formation* consists of gray to red argillaceous limestone and calcareous shale. In Kelly Creek it is approximately 30 feet thick. The *Laurel limestone* in Kelly Creek consists of only 13 feet, is uniformly bedded, gray to pinkish-gray limestone containing scattered particles of red calcite.

### MISSISSIPPIAN SYSTEM

*Kinderhook series.*—The fissile, black, carbonaceous *Chattanooga shale* unconformably overlies the Osgood, Laurel, or Brassfield formations where they are preserved. Elsewhere it rests on the topmost member of the Richmond group. The basal unit of the Chattanooga shale is the dark-gray, phosphatic Hardin sandstone member.

*Osage series.*—The *Fort Payne chert*, a series of residual siliceous siltstone and chert, overlies the Chattanooga shale and is the resistant formation that forms the relatively flat erosional level of the region—the Highland Rim.

## CRITICAL EXPOSURES OF THE PULASKI CHANNEL AND THE SEDIMENTS DEPOSITED IN IT

Locality number (locations indicated on Plates 1 and 2)

1. Conglomeratic sandstone, apparently 6 feet thick, is exposed west of U. S. Highway No. 31. The base of this unit is not exposed here. This is the upper part of the sandstone, and it is finer in texture than is common.

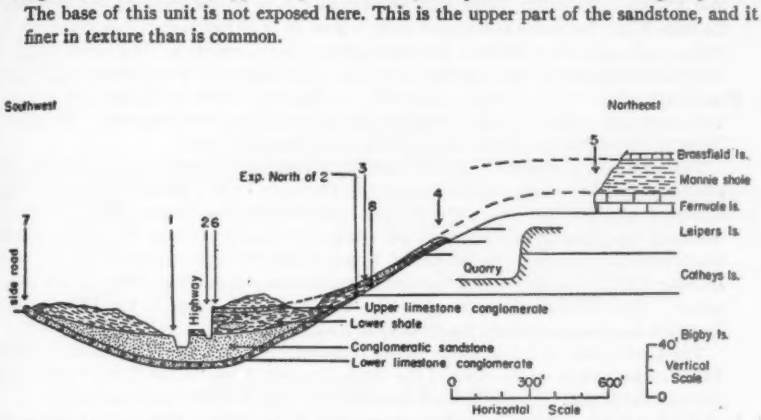


FIGURE 4.—*Vertical cross section across the Pulaski Channel 1¼ miles south of Pulaski and between Localities 5 and 7*

The line of this section is shown on Plate 1

2. Conglomeratic sandstone is also exposed on the east side of the highway a short distance north of Locality 1. Only a few hundred feet north of Locality 2 the Bigby limestone is exposed on the same side of the highway at highway level. The Conglomeratic sandstone, however, wedges out up the slope of the channel between Locality 2 and the exposure of Bigby so that the younger Lower shale rests upon the Bigby.
3. By following the bed of the small branch northeastward from a few feet north of Locality 2 toward Locality 3, an interesting cross section can be studied (Fig. 4). Within a few hundred feet of the highway the Conglomeratic sandstone rises with the gradient of the branch nearly 20 feet above its level at Locality 1. Thise rise is believed to represent a meniscuslike initial dip up the slope of the channel. Occasional beds of Lower limestone conglomerate containing more or less angular blocks of Catheys and Leipers limestones up to a foot in diameter are exposed near the base of the Conglomeratic sandstone. Continuing up the bed of the branch, walk to the left from the branch to an old road. Turn right on this road, following it past the quarry on the left in Catheys and Leipers limestones. A short distance beyond the quarry and in the old road is an exposure (loc. 3) of about 6 inches of Upper limestone conglomerate containing small rounded pebbles of older underlying limestone and quartz grains. It contains Fernvale fossils. This bed of limestone conglomerate is also exposed on both sides of the road. In the road it is underlain by about 3 feet of Lower shale that rests upon the underlying Catheys limestone.
4. Leave the road at Locality 3 and follow the bed of limestone conglomerate to the left up a small-scale dip slope on the Upper limestone conglomerate that rises about 20 feet above its position in the road. Above the south side of the quarry and at the top of the small dip-slope are excellent exposures of Upper limestone conglomerate in a matrix of Fernvale limestone. The conglomerate contains well-rounded pebbles and cobbles of older underlying limestone and some quartz grains. Interbedded with and essentially forming the upper part of the limestone conglomerate are beds of essentially normal Fernvale limestone containing Fernvale fossils

(Pl. 1, loc. 4). Truncation of Leipers and Catheys limestone is well shown just south of the quarry.

5. Northeastward from Locality 4 the drainage divide behind the quarry may be followed toward the prominent hill capped by Mississippian, on the lower slopes of which normal Fernvale limestone and Mannie shale occur. This exposure, Locality 5, is several hundred feet northeast of Locality 4 and the Fernvale limestone there is from 20 to 30 feet higher than the Upper limestone conglomerate in a matrix of Fernvale limestone is at Locality 4. It is believed that this Fernvale limestone was once continuous with the Upper limestone conglomerate at Locality 4.
6. Road intersection. On the southwest side of U. S. Highway No. 31 Lower shale is well exposed both north and south of the side road that leaves the highway southwestward. Several large lenses of fine-grained sandstone occur within the Lower shale here.

   The cut on the northwest side of the highway at Locality 6 includes about 20 feet of Lower shale between the Conglomeratic sandstone at Localities 1 and 2 and the thin bed of Upper limestone conglomerate above the cut just opposite the road interesection. This Upper limestone conglomerate contains sand and small granules of limestone. It contains Fernvale fossils and is correlated with the bed of Upper limestone conglomerate in the old road at Locality 3 and at Locality 4, although lack of continuous exposures prevents tracing them together. The Upper limestone conglomerate at Locality 3 is about 25 feet higher than the limestone conglomerate above the highway at Locality 6.

   The Lower shale, which is about 20 feet thick between Localities 1 and 2 and Locality 6 thins northeastward up the slope of the channel to only 3 feet at the Upper limestone conglomerate at Locality 3, and thins out completely between Locality 3 and Locality 4.
7. In the railroad cut, which is crossed by an overpass, is an excellent exposure of Conglomeratic sandstone. The best vertical exposure of the Lower limestone conglomerate is at the base of the bluff which begins at the northwestern corner of the cut and continues to Richland Creek. There is a good exposure of Lower limestone conglomerate containing pebbles, cobbles, and boulders of Leipers and Catheys limestones in a greenish calcareous matrix containing many grains and granules of quartz in a small cut on the side road about halfway between Localities 6 and 7.
8. In an old lane below a garage north of the house on the east side of the highway is an exposure of Lower shale with a strong initial dip cutting across beds near the base of the Catheys limestone Conglomeratic sandstone is exposed in the bed of the branch a short distance to the northeast.
9. Glades developed on the Lower shale.
10. Conglomeratic sandstone exposed in the bed of a creek.
11. Good exposures of Conglomeratic sandstone in the bed of a creek and along both sides of the creek.
12. Southeastward from exposures in the bed of the creek and up a small but prominent spur, the Conglomeratic sandstone is well exposed. An interesting part of this exposure is the strong initial dip of the southwestern side of the spur. Above the Conglomeratic sandstone there are extensive glades developed on the Shale unit, both the Lower shale and the Upper shale. Near the middle of the Shale unit, and separating the Lower shale from the Upper shale, is a 2 to 3 inch bed of Upper limestone conglomerate. The glades occur on both sides of the drainage divide and in the low saddle (Loc. 12) to the east through which the T.V.A. transmission line passes.
13. A section of normal Richmond, that is, Fernvale limestone, Mannie shale, and Sandstone member, occurs in the southwestern part of the spur northeast of Tarpley. The Sandstone member of the Mannie is well exposed and may be followed northeastward almost to Locality 12. This exposure is particularly important as the member and the overlying Brassfield limestone may be walked about continuously across the deeper part of the channel-filling to their positions above the normal regional sequence (Fig. 5).
14. A terrace-like spur protrudes southeastward from the main northeast-southwest ridge and is supported by the Conglomeratic sandstone.
15. An elongated hill capped by Conglomeratic sandstone. The best exposure of the sandstone is at the northern end of the hill.

16. Elongated twin-hills approximately 1 mile long that are capped by the Shale unit. On the south side of the northern hill two layers of Upper limestone conglomerate occur near the top of the hill indicated by Locality 16 on Plate 1. The lower one is about 1 foot thick and overlies at least 40 feet of Lower shale. The next layer is lenticular in occurrence, 6 inches being its

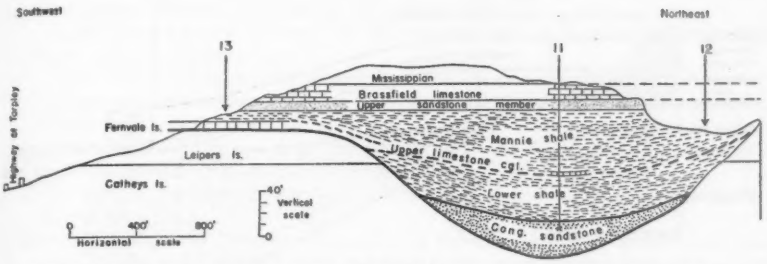


FIGURE 5.—*Vertical cross section across the Pulaski Channel, northeastward from Tarpley to Locality 12*

The line of this section is shown on Plate 1

maximum thickness, 15 feet above the lower one, and about 15 feet below the top of the knob, which is capped by a possible third layer of Upper limestone conglomerate that may or may not be in place. The beds of Upper limestone conglomerate are fossiliferous, and contain well preserved specimens of *Bythopora delicatula*, *Rhombotrypa subquadrata* ?, *Dalmanella meeki*, *Ceratopsis robusta*, and other species of Richmond ostracodes. The presence of these fragile fossils here demonstrates that the Richmond species lived in the submerged estuary and could not have been reworked from older beds.

17. Conglomeratic sandstone in bed of branch.
18. Lower shale overlying Catheys limestone on the west side of the highway. West of this exposure is a low glade-hill on which occurs a bed of Upper limestone conglomerate that is 2 to 3 inches thick.
19. Good exposures of Conglomeratic sandstone along a small branch. The direction of dip of the cross-bedding is well shown here.
20. About 8 feet of Conglomeratic sandstone is exposed in Buchanan Creek and is overlain by 70 feet of greenish-gray shale of the Shale unit with sandy lenses and thin sandy layers. No beds of Upper limestone conglomerate were found in the shale and it is probably composed of both Lower shale and Upper shale.
21. Fair exposure of shale extending southeastward beneath normal Brassfield limestone.
22. Middle of an elongate ridge about 1 mile long capped by Conglomeratic sandstone, the best exposure of which occurs at the northern end of the ridge.
23. Ridge capped by Conglomeratic sandstone.
24. Good exposure of Conglomeratic sandstone with particularly good cross-bedding.
25. In a new highway cut about one-half mile south of Elkton, the lower part of the Conglomeratic sandstone is well exposed resting upon limestone beds near the base of the Catheys limestone. The lower 1 to 3 feet of the sandstone contain small to large, angular to round fragments of limestone, chert, etc. This and Locality 7 B are the best exposed contacts between the Channel-filling sediments and the older underlying limestone.
26. Conglomeratic sandstone is exposed in the bed of the creek and is overlain by about 100 feet of Shale unit in which no Upper limestone conglomerate was found. The hill north of the exposure in the creek is capped by the Sandstone member of the Mannie and fragments of Brassfield limestone. Marine fossils occur near the base of the Sandstone member.
27. Conglomeratic sandstone in creek.

28. Conglomeratic sandstone is well exposed in the railroad cut. The sandstone is also exposed in each major southward draining branch of Sinking Creek between Localities 27 and 28.
29. Along the road that follows the western headwater tributary of the Reed Branch are exposed, in order, Mississippian beds, Brassfield chert, Mannie shale, Fernvale limestone (only 2 to 4

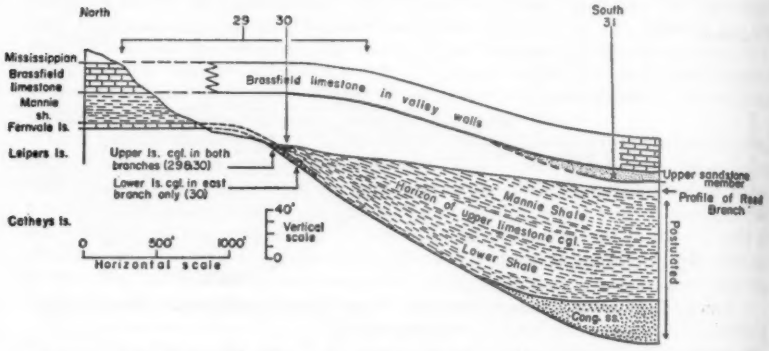


FIGURE 6.—*Vertical cross section across the Pulaski Channel along Reed Branch between Localities 29 and 31*

The line of this section is shown on Plate 2

feet thick), and Leipers limestone, which is exposed in a small inlier between this normal sequence and the channel-filling sediments to the north. At the northern end of this inlier (Pl. 2, loc. 29), the Leipers is overlain by about 3 feet of fine-grained sandstone (not shown in Figure 6) and an excellent exposure of Upper limestone conglomerate that extends at least 100 feet down the creek. This lens of conglomerate wedges out with such surprising abruptness that it suggests a mass of debris of pebbles and cobbles of older underlying limestone mixed with Fernvale ooze that slipped down the slope of the channel out upon the Lower shale. A foot or so of Lower shale is exposed beneath the thinned edge of the conglomerate.

The Upper limestone conglomerate, in turn, is overlain by Upper shale dipping downstream and toward the deepest part of the channel at a rate equal to or slightly greater than the gradient of the creek. The shale is exposed in the bed of the creek past the junction of the eastern headwater tributary of Reed Branch northward toward Locality 31.

30. Along the eastern headwater tributary of Reed Branch practically the same sequence of exposures is found except that the fine-grained sandstone (Loc. 29) is absent and a small exposure of Lower limestone conglomerate is present between two exposures of Upper limestone conglomerate. Undoubtedly, the latter has been raised several feet by structural means, but even so, it is believed to represent Lower limestone conglomerate that was accumulated fairly high upon the channel slope contemporaneous with the younger part of the Lower shale. Its postulated normal position is indicated in Figure 6 in which the exposures of Localities 29, 30, and 31 are shown.
31. Exposures of the Sandstone member of the Mannie shale along Reed Branch.
32. Good exposure of almost 10 feet of Upper limestone conglomerate overlain by a covered interval (Mannie shale?), and loose fragments of the Sandstone member and Brassfield chert and limestone. The occurrence of Leipers limestone relatively high on three sides of this exposure suggests that this perched patch of Upper limestone conglomerate accumulated in a short tributary with a steep gradient rather than in the master channel.
33. Conglomeratic sandstone is well exposed in the bed of a creek opposite the spring on the west side of the road. About 25 feet above the top of the Conglomeratic sandstone (and above the

spring) is a 7 foot bed of Upper limestone conglomerate. This conglomerate contains rounded pebbles and cobbles of Leipers and Catheys limestone and granules and grains of quartz imbedded in a matrix of Fernvale limestone. The matrix and certain of the beds resemble normal Fernvale limestone very closely. This is the only known occurrence of thick massive Upper limestone conglomerate near the middle of the channel-fill. The conglomerate is overlain by about 100 feet of the Shale unit with occasional sandstone layers, the Sandstone member of the Mannie shale, Brassfeld limestone, and Mississippian strata.

In the bed of a branch in a barnyard about 100 feet north of the spring is an exposure of Upper limestone conglomerate.

South of the Conglomerate sandstone in the bed of the creek the channel wall, composed of Catheys and Leipers limestone, is estimated to rise 60 feet within a horizontal distance of about 350 feet.

A good exposure of normal regional Fernvale limestone and Mannie shale may be seen in the quarry on the east side of the road about 3 miles south of Coldwater.

34. About 35 feet of massive Conglomeratic sandstone that shows the nature and direction of dip of the cross-bedding.

35. Overlying normal older limestone and continuing northward up the spur to the prominent rounded, bald-topped hill, the following sequence occurs: 3 feet of Lower Conglomeratic sandstone, 15 feet of Conglomeratic sandstone, about 100 feet of the Shale unit with occasional layer of sandstone, and Brassfield limestone. The top of the Conglomeratic sandstone here can be walked northward to the sandstone exposed at Locality 34.

36. Conglomeratic sandstone exposed in the bed of the Creek. An excellent prospective (Fig. 7) of the channel wall may be gotten here by walking northward up the creek and turning eastward toward Locality 37 up the second side-branch on the right, or east. The Conglomeratic sandstone rises up the branch with an initial dip locally as high as 15°. Above the meniscus-like edge of the sandstone is an exposure of Lower limestone conglomerate plastered high on the slope of the channel and upon the older underlying limestone into which the channel was cut.

37. This station is located in the saddle between the prominent rounded glade-hill south of the saddle and the spur to the north. The Shale unit is well exposed on all sides of the glade-hill. Upper limestone conglomerate occurs in two parts here, located in reference to the saddle, as follows:

1. Two large blocks lying loose on the shale in the saddle.
2. Ledge forming the eastern edge of the saddle over-looking the valley to the east. Some of the conglomerate in this exposure is normal Fernvale limestone.

Normal Richmond occurs about 1500 feet north of the saddle although it is practically all covered except the Sandstone member of the Mannie. The entire distance of 1500 feet is along a narrow ridge capped with Leipers limestone.

38. Spur capped by Conglomeratic sandstone.

39. Good exposures of Conglomeratic sandstone at the end of a prominent hill. Here the sandstone is about 35 feet thick and overlies 3 feet of Lower limestone conglomerate.

40. Conglomeratic sandstone is exposed in the bed of the branch and is overlain by the Shale unit on the west side of the side road. This sandstone is also exposed in all the major creeks between Locality 39 and Locality 40.

41. Exposures of Upper limestone conglomerate and shale on the side of a ridge represent accumulation in a short tributary to the master channel as they are perched considerably above the level of the channel. The tributary was approximately 1500 feet in maximum width at this locality. Normal Fernvale limestone is exposed north of this subsidiary channel.

42. An unusual exposure of the Conglomeratic sandstone occurs in the face of the hill where the channel-filling sediments disappear beneath younger strata. Only about 25 feet of the sandstone are exposed here, but it consists of 3 beds of Lower limestone conglomerate, each averaging about 4 feet thick, that alternate with beds of Conglomeratic sandstone of about equal thickness. This entire exposure has a marked initial dip.

Five feet below the top of the exposure and just above the top bed of the Lower limestone conglomerate there is a 6 inch bed of normal Fernvale limestone. The calcareous matrix of all of the Lower limestone conglomerate everywhere is strongly suggestive lithologically of the

Fernvale limestone, but this is the only known occurrence of undoubted Fernvale limestone interbedded with the Lower limestone conglomerate and the Conglomeratic sandstone It supports the statement that the Conglomeratic sandstone and Lower shale with their closely associated Lower limestone conglomerate are all of early Fernvale age" and are equivalent to the normal regional 'Fernvale sequence" to the southeast in northeastern Alabama.

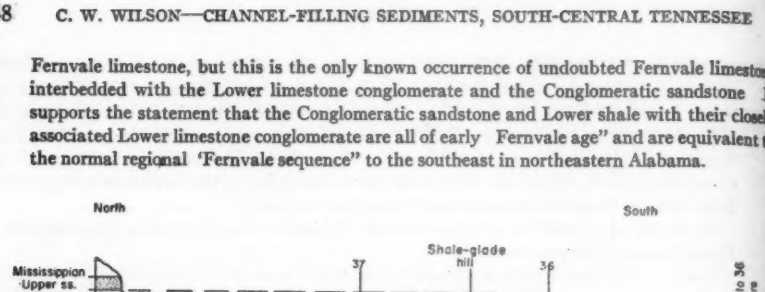


FIGURE 7.—*Vertical cross section across the Pulaski Channel 2 miles southeast of Coldwater, between Localities 36 and 37*

The line of this section is shown on Plate 2

## FAYETTEVILLE CHANNEL

### INTRODUCTORY STATEMENT

In the central part of Lincoln County four small areas (Fig. 1) of thick clastic Richmond sediments are identical with those described in the Pulaski Channel. If these four areas were the only occurrences of this type of sediments, they would be inadequate for a convincing interpretation of origin, but fortunately, the well-preserved channel-filling sediments of the Pulaski Channel supply the data necessary to interpret them. The postulation that these four areas are remnants of the fill of a channel similar to the Pulaski Channel is based entirely upon the similarity of sediments.

The four areas located near Petersburg, Howell, Fayetteville, and Kelso, occur in an essentially straight line following the structural grain of the region as in the case of the Pulaski Channel. The total length of the postulated Fayetteville Channel is 16 miles, of which the four areas of preserved channel-filling sediments make up only about 2 miles.

It is believed that the cutting of the Fayetteville Channel and the filling of this channel with thick clastic Richmond sediments were similar to that of the Pulaski Channel but with two notable differences:

(1) The Calcareous mudstone member of the Fernvale formation of the normal regional sequence occurs below the Conglomeratic sandstone of the thick clastic sequence at Howell, Fayetteville, and Kelso. The nature of the exposure at Petersburg makes it impossible to determine the presence or absence of this member there. Near these localities this member is restricted to the channel, indicating that the

sea in which this, the earliest Richmond unit in south-central Tennessee, was deposited extended at least 12 miles up the channel as an estuary and that the occurrences at Howell, Fayetteville, and Kelso are of estuarine origin but are contemporaneous with the marine Calcareous mudstone member of southeastern Lincoln County, the southern half of Franklin County, and the region to the southeast.

(2) The Fayetteville Channel coincides with the crater of cryptovolcanic or meteoric origin at Howell. At the present site of Howell along the Fayetteville Channel an event occurred during the post-Maysville and pre-Richmond interval of erosion that was to play an indirect role in the development of that stream channel. At that time an explosion (Born and Wilson, 1939) blew out a crater with minimum depth of 100 feet, maximum depth of 200 feet, and diameter of 1 mile. The limestone debris, piled up around the rim of this crater, was ultimately removed. The crater walls were graded by erosion. During this interval the crater was undoubtedly partially or completely filled with fresh water. Lake sediments probably were deposited within the crater, but, as the lower part of the center of the crater has been removed by relatively recent erosion, no trace of the lake sediments remains. The flooding and overflowing of this crater established a normal line of drainage southeastward down the slope of the region and parallel to the structural grain. On the upper, or northwest, rim of the crater, headward erosion started to work, gradually extending the line of drainage northwestward. Further deepening of this line of drainage, located in part by the Howell crater, reduced the level of the lake, possibly removed some of the lake sediments, and completed the development of the Fayetteville Channel. This channel would have been similar in appearance to the Pulaski Channel except for the presence of this circular lake about 1 mile in diameter in sharp contrast to the channel's average width of a few thousand feet.

### CRITICAL EXPOSURES OF THE FAYETTEVILLE CHANNEL AND THE SEDIMENTS DEPOSITED IN IT

1. Petersburg. About 1 mile south of Petersburg (Fig. 8) there is a pile of Conglomeratic sandstone about the size of a small house, very similar lithologically to that of the Pulaski Channel but more ferruginous, probably due to the nature of the exposure. The sandstone occurs in a jumbled mass of blocks upon the outcrop belt of the Cannon limestone. Although the blocks are not strictly in place, it is believed that they have not been lowered many feet. The more resistant nature of the sandstone permitted it to be not only preserved but also etched into relief. Normal regional Richmond occurs in the higher hills in the vicinity of Petersburg.
2. Howell. The thick clastic Richmond sediments occur in a north-south range of hills about three-fourths of a mile east of Howell (Fig. 9). In relationship to the Howell crater, the sediments remaining today are those that were deposited high upon the walls of the crater-channel, those deposited in the deeper part have been entirely eroded. The Richmond strata dip with varying degrees of steepness and of strike, resulting either from post-Richmond renewal of the force that had earlier formed the crater or from differential settling. The preservation of such a small percentage of the sediments that had originally filled the crater-channel and the structural disturbance of these beds make a complete and convincing interpretation of all details difficult.

   The general relationships here are seen to best advantage by walking up the bed of the small branch which crosses the secondary road north of Howell (Fig. 9, loc. 1), to its head and then turning southeastward to Locality 2 (Fig. 9), where the road eastward from Howell crosses the prominent topographic saddle.

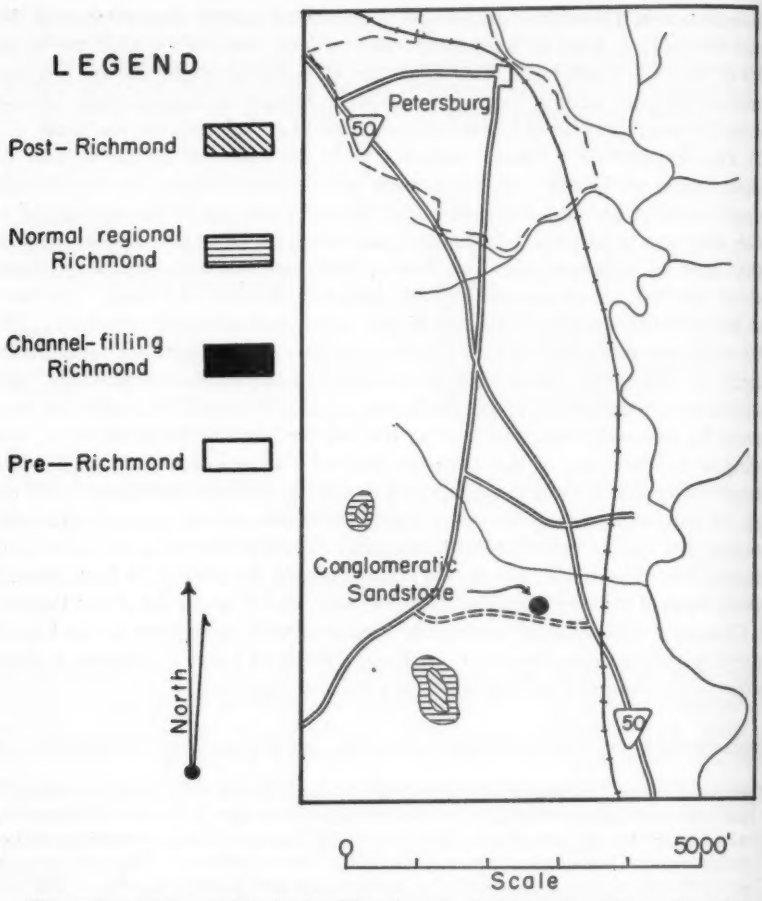


FIGURE 8.—*Index map of the vicinity of Petersburg showing the location of the small hill of Conglomeratic sandstone*

The Calcareous mudstone member of the Fernvale formation, which was of estuarine origin here, is restricted to the deeper part of the channel and crater (Fig. 9). It is the usual green, calcareous siltstone and mudstone averaging about 10 feet thick.

The Conglomeratic sandstone (Fig. 9, loc. 3, 4, 5), is similar in thickness, lithological characteristics, and texture to that of the Pulaski Channel. At Locality 4, the lowermost 12 feet of the sandstone is well exposed as an almost plano-convex lens about 75 feet wide. Residual Brassfield chert and fossils occur above the lens, indicating local removal of all Richmond sediments deposited after the Conglomeratic sandstone. At Locality 3 the sandstone forms the saddle between two hills capped with Mississippian strata. The lens is almost plano-convex in cross-section and is about 650 feet wide. The wedging ends of the lens are about 15 feet lower than the middle. Locality, it is overlain by Mississippian strata. A short distance to the south the Mississippian overlies the Lower shale, which is post-Conglomeratic sandstone in age. Several exposures of the sandstone occur along the north side of the ridge (Fig. 9, loc. 5).

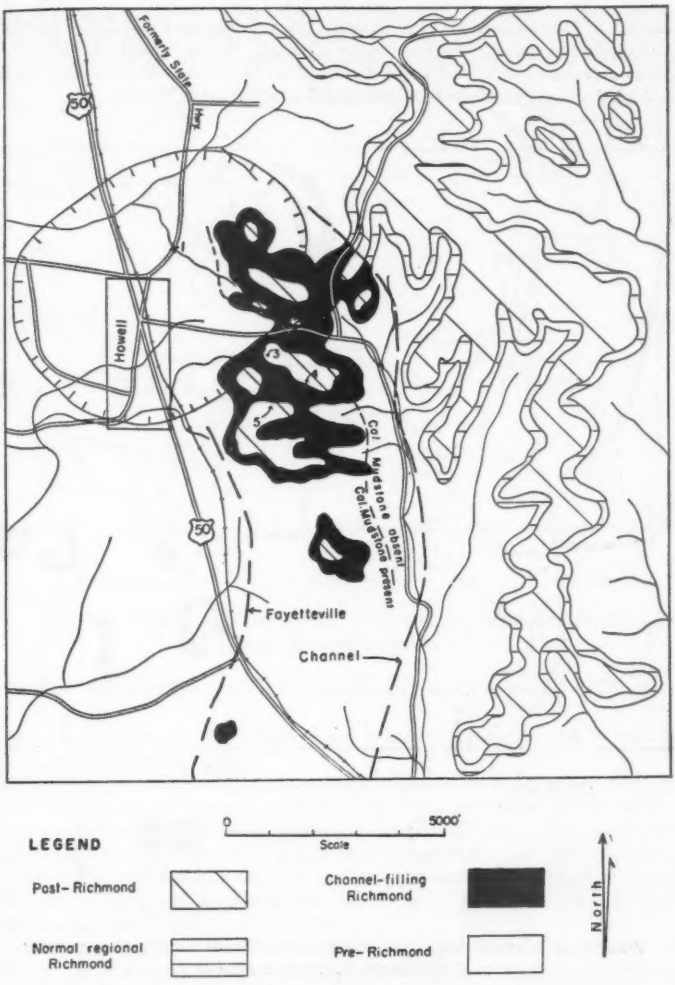


FIGURE 9.—*Geologic map of the vicinity of Howell showing the occurrence of thick clastic Richmond sediments*

The Howell cryptovolcanic structure is indicated by the circular depression contour line

The outstanding anomaly of the Conglomeratic sandstone here is that the exposures occur at a much higher elevation in reference to the Fernvale limestone exposed to the north than they normally should. Lack of exposures at critical points prevent tracing the Fernvale limestone and the Conglomeratic sandstone together in a vertical exposure where the anomaly could be studied. It is the writer's belief, however, that the area in which the Conglomeratic sandstone is present was uplifted by local post-Richmond and pre-Brassfield deformation acting in such a way as to raise the sandstone to the approximate level of the Fernvale limestone to

the north. This belief is based to a large degree upon the assumption of similar age and history of the Pulaski Channel and the Fayetteville Channel. In the Pulaski Channel the Fernvale limestone averages about 40 feet above the top of the Conglomeratic sandstone, and the same general relationship is predicated for channel-filling sediments at Howell.

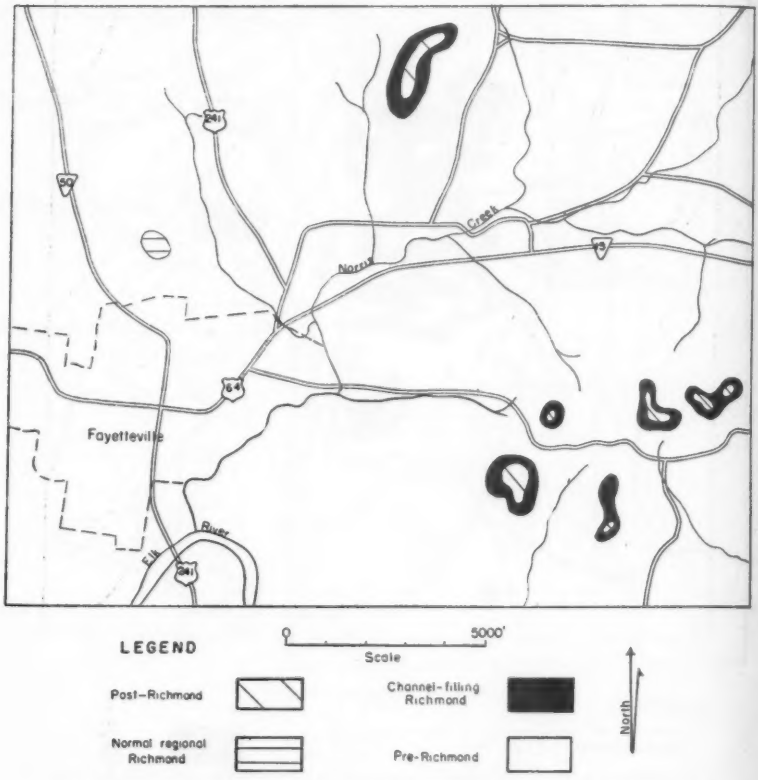


FIGURE 10.—*Geologic map of the vicinity of Fayetteville showing the occurrence of thick clastic Richmond sediments*

The Lower shale is undoubtedly well represented in the Howell area, but due to the poor exposures and the meager preservation it is usually impossible to identify it as a specific unit. Much of the shale in the southern part of the area of preservation is probably Lower shale.

The irregularly bedded, massive, coarsely crystalline, red Fernvale limestone is fairly well exposed, ranging from 10 to 35 feet thick. At the head of the small branch that crosses the road north of Howell at Locality 1, the lower beds of the Fernvale limestone are in part brecciated. The cobbles of Fernvale limestone, however, are imbedded in a matrix of identical Fernvale limestone. There are no fragments of the underlying breccia of Carters, Hermitage, etc. strata in the Fernvale breccia as the only Fernvale limestone preserved today was deposited near the top of the crater and apparently above the mass of rubble. It does overlie a breccia of older limestone, but it is more in the nature of massive shattered limestone, incapable of

serving as a source of pebbles and cobbles. The Fernvale limestone maintains a distinct unity as a veneer covering the underlying shattered limestone. It is believed that the brecciation of the Fernvale limestone is a local phenomenon that occurred shortly after the partial consolida-

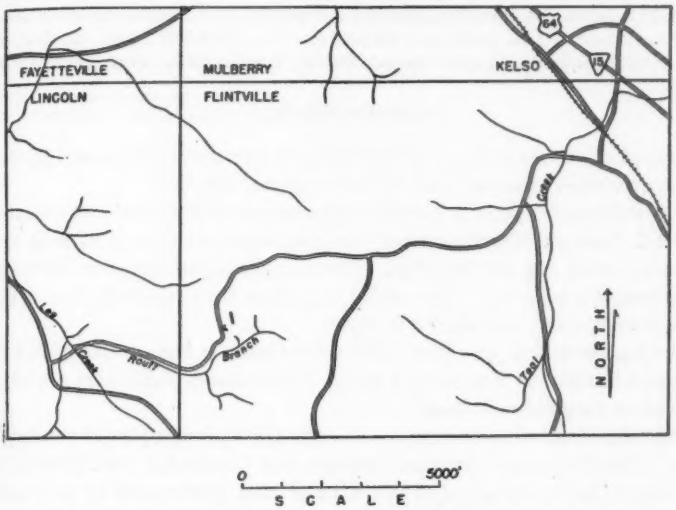


FIGURE 11.—*Index map of the vicinity of Kelso showing the location of thick clastic Richmond sediments*

The names "Fayetteville", etc. refer to the quadrangles (7½') of the Tennessee Valley Authority.

tion of some of the beds of Fernvale limestone, as a result of readjustment in the underlying jumbled breccia of older limestones.

The Mannie shale consists of greenish- to grayish-brown shale and mudstone with occasional layers of siltstone and fine-grained sandstone. The lower part of the Mannie shale contains many coarse grains and granules of quartz which are believed to have been locally derived by the reworking of the upper part of the Conglomeratic sandstone. The presence of a few crinoid stems and brachiopods indicates its marine origin. Its thickness is difficult to measure with accuracy due to lack of continuous exposures and to erratic and variable dips, but it is believed to reach a maximum of 95 feet east of Howell.

The sandstone member of the Mannie shale is exposed at several places north of Locality 2 (Fig. 9). It consists of the characteristic very fine-grained sandstone approximately 10 feet thick. Here it is overlain by Mississippian strata and overlies Mannie shale.

At the time the work was done on the structure at Howell the origin and history of the thick clastic Richmond sediments were not understood. Their age and sequence were recognized but it was then assumed that they were part of a more widespread blanket of sediments. The details of the sedimentary filling of the Pulaski Channel had not been worked out at that time. For these reasons several changes have been made in the original description of the crater-filling sediments (Born and Wilson, 1939). These changes in stratigraphic interpretation are only minor and in no way affect the interpretation of events in the origin of the disturbed area at Howell.

3. Fayetteville. On a hill about 1½ miles northeast of Fayetteville and several hills about 2 miles east of Fayetteville (Fig. 10) there are exposures of the Calcareous mudstone member overlain by Conglomeratic sandstone. The latter is overlain by Mississippian strata. It is believed

that the Calcareous mudstone member and the Conglomeratic sandstone of these hills represent remnants of the basal part of the channel-filling sequence. Normal Fernvale limestone is well exposed capping the hill just north of Fayetteville.

4. Kelso. Near the head of Routt Branch of Lee Creek about 2 miles southwest of Kelso (Fig. 11) several feet of the Calcareous mudstone member underlie the Conglomeratic sandstone. The exposure is located in the road and in the field east of Locality 1 (Fig. 11). At this point the channel-filling sediments disappear beneath the Mississippian strata of the Highland Rim.

## CONCLUSIONS

The record of stream cutting and of filling of the resulting channel during the Ordovician period is believed to be unique in several respects:

(1) A post-Maysville stream flowed southeastward off the southeastern extension of the Ozark dome across south-central Tennessee and cut a channel at least 32 miles long, 1500 feet wide, and 100 feet deep. This requires higher elevation for the region than was formerly believed. The cutting must have been relatively fast as there is evidence of only a very few short tributaries.

(2) The bottom and slopes of the channel were more or less covered with a rubble of pebbles and cobbles of Trenton and Maysville limestones which were reworked and redeposited as conglomeratic lenses.

(3) The Richmond sea, advancing northwestward from that part of the Appalachian geosyncline that is now northeast Alabama, first extended at least 32 miles inland up this channel as an estuary, up which silt and sand were carried by tidal action of the advancing sea. The very low gradient of the channel permitted the flooding to extend so far above its mouth, which was probably many miles southeast of the present site of Coldwater. This is the only recorded case of Lower Paleozoic sediments being carried up-channel in an estuary against the original direction of fresh water down the stream, but the evidence presented by the calcareous matrix of the sandstone, by the direction of dip of the foreset beds of the cross-bedding, and by the presence of similar sediments to the southeast and their absence to the northwest all require this interpretation.

(4) The channel was three-fourths filled with estuarine sediments before the region was flooded and the channel submerged by the sea in which the Fernvale limestone was deposited. A thin limestone conglomerate (Upper) represents the sediments in the channel that are contemporaneous with the Fernvale limestone of the region.

(5) The submerged channel was filled by silt that became the Mannie, or Upper, shale so that the younger part of the Mannie shale and the Sandstone member of the Mannie shale—the uppermost unit of the Richmond group in Central Tennessee—were deposited uniformly over the top of the filled channel.

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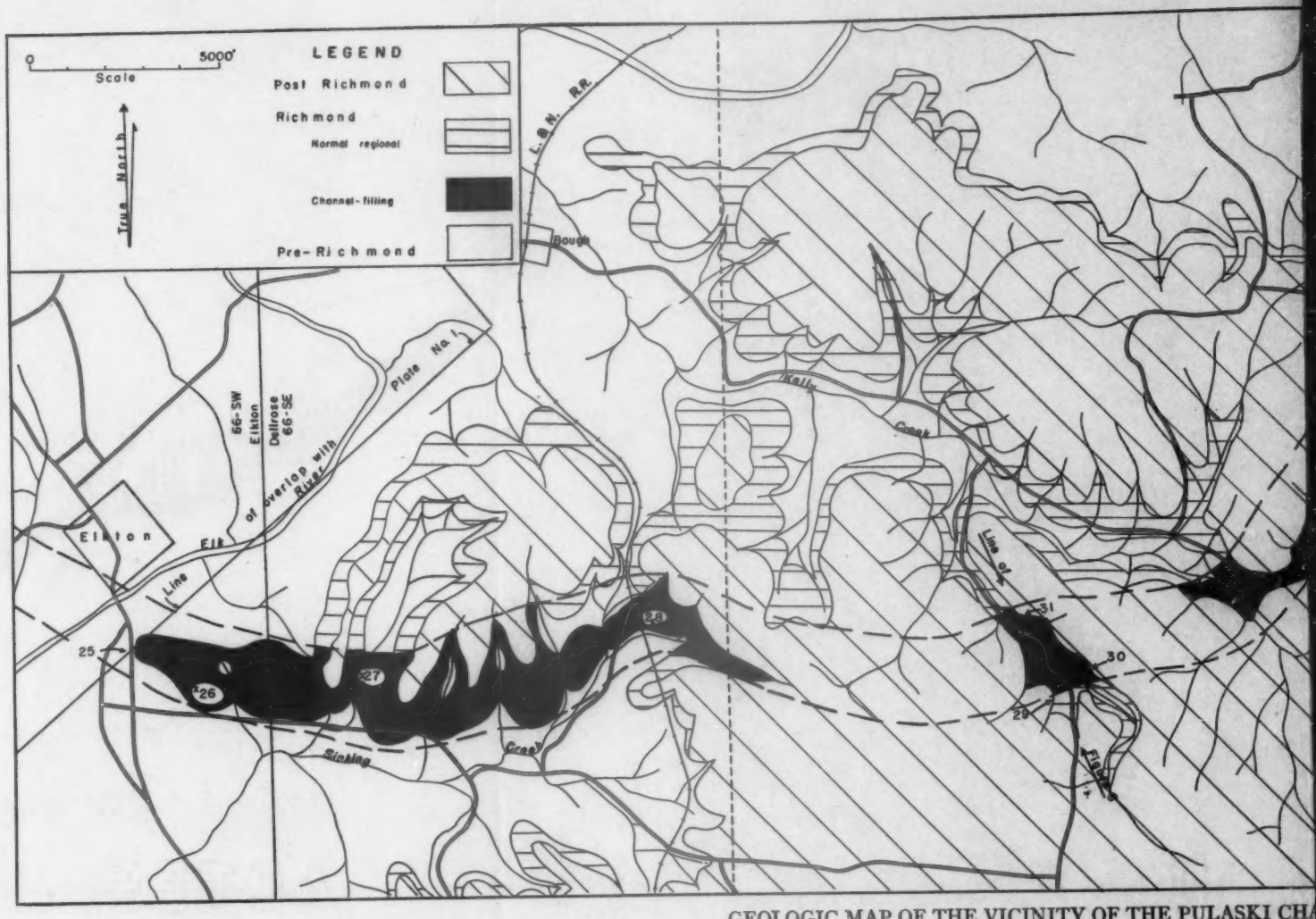


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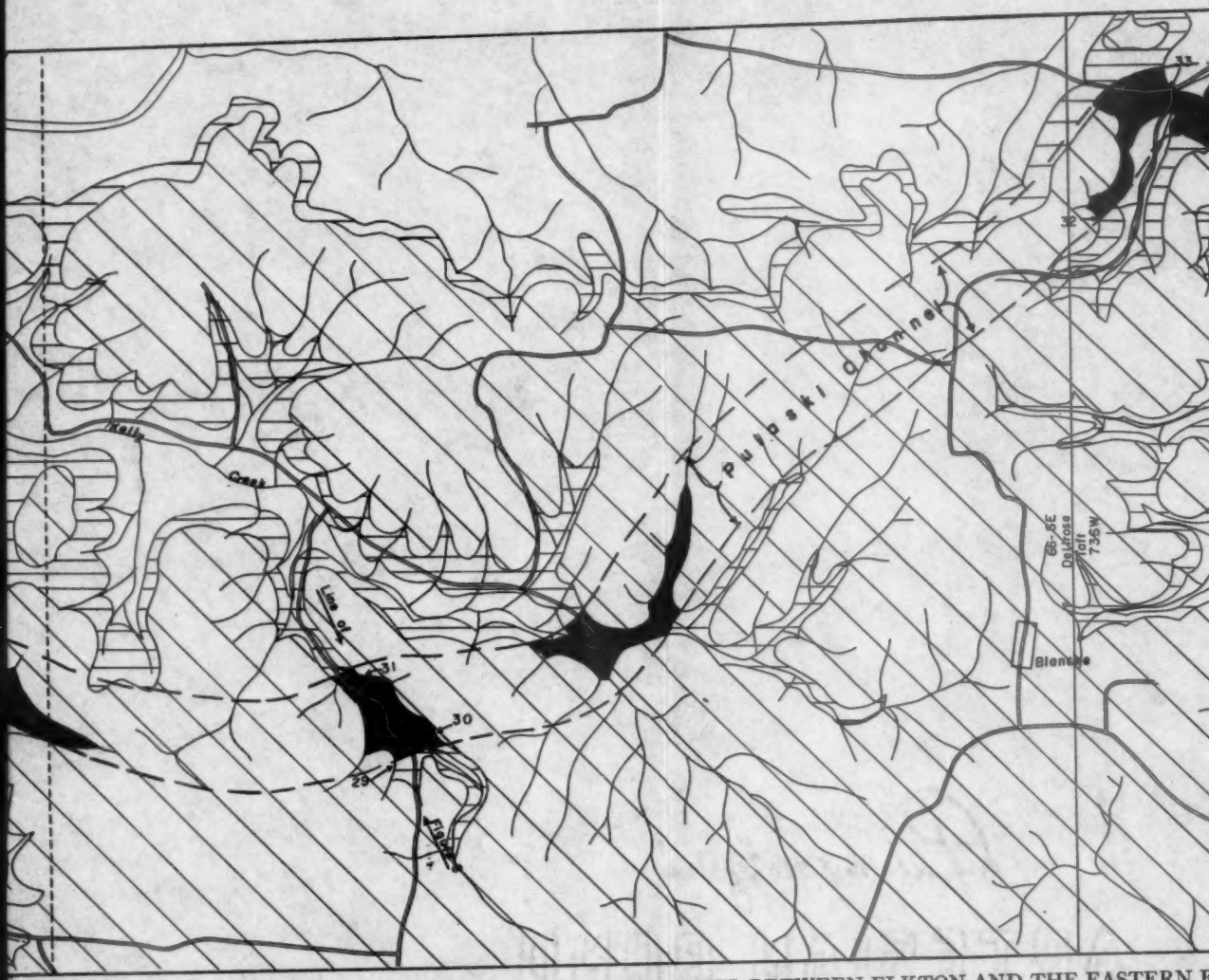


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# APPLICATION OF BRIANCHON'S THEOREM TO CONSTRUCTION OF GEOLOGIC PROFILES

BY JOHN B. MERTIE, JR.

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## ABSTRACT

Brianchon's theorem states that the three diagonals joining opposite vertices of a hexagon circumscribed about a conic are concurrent. A corollary of this theorem applies to a pentagon so that the points of tangency of an inscribed conic may be located.

Any five non-concurrent straight lines in a plane, no three of which are parallel, will ordinarily form some kind of a pentagon; and if considered as tangents to a conic, they will define its shape and position. If these five lines are also normals to the traces of parallel stratigraphic surfaces having a constant strike, the derived conic

may be regarded as an evolute, from which a set of involutes can be drawn that will constitute a structural profile. A method is thus afforded for constructing a profile normal to the strike of the rocks, and for measuring stratigraphic thickness, by the utilization of five observations of dip along a suitable linear traverse. Graphical methods are also given for constructing a parabolic evolute from four observations and a circular evolute from three observations. Additional points on the conic evolutes are obtained by the application of Pascal's theorem.

A mathematical analysis is presented of the relationship between a conic and five of its tangents; and the conic evolute, rather than its involutes. is recommended as a satisfactory record of the structure of a parallel fold. To obtain the equation of this evolute, the equations of the five tangents are first derived, using trilinear coordinates. Thereafter the tangential and trilinear equations of the general conic are deduced. Criteria are given for classifying the conic as a hyperbola, ellipse, parabola, or circle.

These graphical and analytical methods are adaptations of the general method of evolute and involutes. They are offered, not as substitutes for the general method, but as quicker, though somewhat less accurate, means of obtaining similar results, where structural conditions justify their use.

## INTRODUCTION

Two methods are known for the mechanical construction of geologic profiles of parallel folds. The older and better known of these is the method of concentric arcs, described first by Hewett (1920, p. 367–385), and extensively used by Busk (1929). The second is the general method of evolute and involutes, proposed by the writer (Mertie, 1940, p. 1107–1134), and further developed in a recent paper (Mertie, 1947, p. 779–802). The Hewett method is restricted to the utilization of structural observations in pairs, whereas the general method of evolute and involutes may utilize any number of structural observations that lie within or close to the profile plane.

This paper proposes new methods by means of which five, four, or three observations of dip lying within or close to a profile plane may be used to construct conic evolutes, from which involutes may be drawn that will represent approximately the traces of parallel stratigraphic surfaces. The methods that require five or four observations are based upon a corollary of Brianchon's theorem. The method that utilizes three observations is based upon the theorem that three tangents to a circle define its locus. These techniques, all of which yield conic evolutes, are designated collectively as the special method of evolute and involutes, to distinguish them from the general method of evolute and involutes, which yields evolutes that are higher plane curves.

Profiles through parallel folds are commonly drawn to show the true curvature of stratigraphic surfaces and the stratigraphic thickness. For the attainment of these two objectives, it is necessary that the profile shall be normal both to the stratigraphic surfaces and to the axis of the fold. Therefore, for all kinds of parallel folds, the profile may lie in a plane, or in a cylindrical surface, or in a doubly curved surface. The traces of the stratigraphic surfaces in all such profiles will be parallel lines, but such traces or involutes may be either plane curves in a plane, or space curves in a curved surface. In practical geological applications, only plane profiles are constructed.

Cylindrical folds constitute the principal type of structure in which the method of

evolute and involutes can be used. In this designation, no restriction is placed on the geometric form of the cylindrical surfaces, beyond the condition that they be parallel. Cylindrical folds may be horizontal or nonhorizontal. If a parallel cylindrical fold has a constant strike throughout some specified interval, its axis within that interval is horizontal; and a linear traverse normal to the strike will lie in a vertical plane that is normal both to the axis of the fold and to its stratigraphic surfaces. The traverse, however, may be made oblique to the strike, as the sites of the observations can be projected along the strike into the vertical profile plane. If the axis of a cylindrical fold in nonhorizontal, the required profile will lie in an inclined plane normal to the axis of the fold. In a linear traverse within this plane, the strike will change from point to point, but this will cause no difficulty provided the change is caused only by the pitch of the fold. The recorded dips, however, are referred to a vertical plane; and if they are to be used in an inclined profile, they must be projected through an angle equal to the pitch of the fold, if that is determinable. The traverse does not necessarily have to be made in the plane of the profile; but if it is not, the sites of the observations will have to be projected into the profile plane along lines parallel to the axis of the fold. With these limitations, inclined plane profiles may also be made.

Quaquaversal folds constitute a second type of structure that may yield plane profiles, wherein the traces of parallel stratigraphic surfaces are involutes, though more commonly the profiles will lie in curved surfaces and the involutes will be space curves. The necessary restriction is that the quaquaversal fold shall have one or more planes of symmetry, and that the traverse shall be made in such a plane.

A vertical plane profile showing the traces of parallel stratigraphic surfaces as involutes is justified mathematically only where the strike is constant throughout the length of a traverse. But in actual practice the optimum conditions, though lacking, may be so closely approximated that the method is warranted. A tilted plane profile, for example, may depart so little from the vertical that a delineation of involutes in a vertical profile may result only in slight error. Similarly the substitution of a plane profile for a curved one may cause only a slight distortion of the involutes. Therefore the method of evolute and involutes, restricted to specified intervals, has many applications besides those wherein it is mathematically exact.

## BRIANCHON'S THEOREM

Brianchon's theorem states that the three diagonals joining opposite sides of a hexagon circumscribed about a conic will meet in a point. In Figure 1, let ABCDEF be a convex hexagon circumscribed about an ellipse. The three diagonals from the six vertices will meet in the point O. Now allow the internal angle formed by any two sides, say ABC, to approach 180°, so that the sides AB and BC approach collinearity. At the limit, the broken line ABC will become the straight line AC, and the point B will become one of the points of tangency of a circumscribed pentagon.

A convex pentagon, circumscribed about an ellipse, is shown in Figure 2. Draw the diagonals AD and CF, thus locating the intersection O, as in Brianchon's theorem. Then draw the line EO, and extend it to meet the side AC. The intersection, K, is

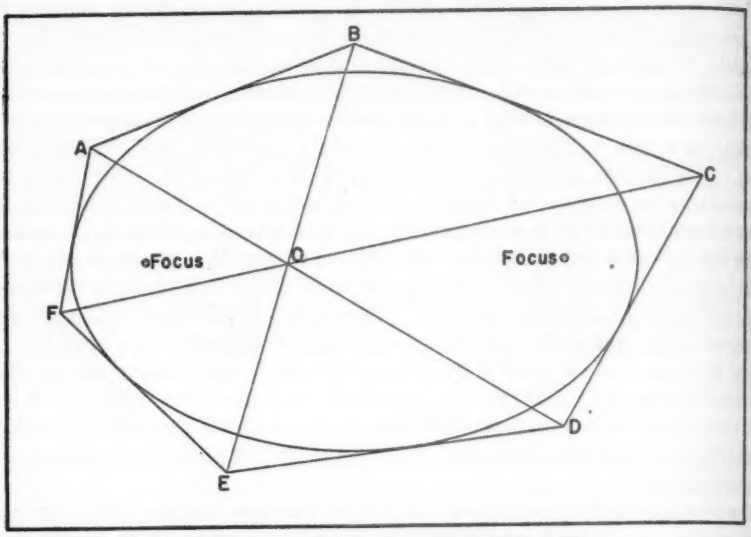


FIGURE 1.—*Convex hexagon circumscribed about an ellipse*

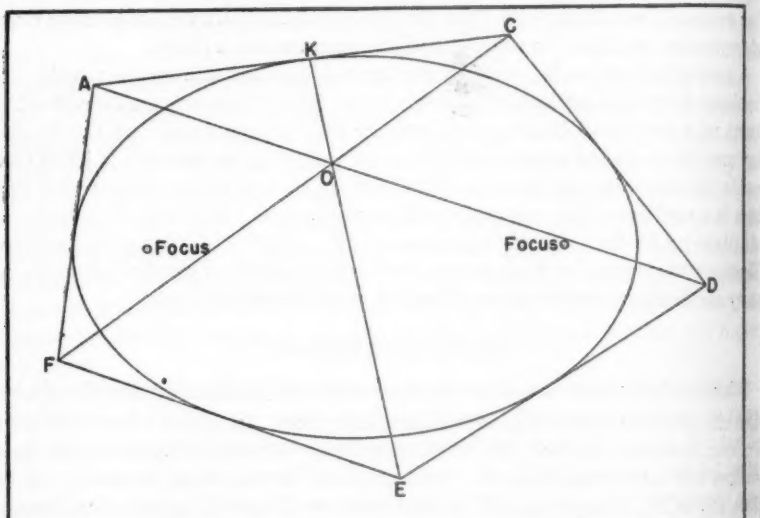


FIGURE 2.—*Convex pentagon circumscribed about an ellipse*

the point of tangency of the line AC with the ellipse. By means of four similar constructions, the other points of tangency of the pentagon will be obtained.

The demonstrations of Figures 1 and 2 were made for an ellipse because the con-

structional lines are clearer; but Brianchon's theorem, and its corollary for a circumscribed pentagon, hold for all nondegenerate conics. In the geologic application of this theorem, however, the five-sided figures may have re-entrant angles, or their sides may actually intersect one another. Hence convex, concave, proper, and improper pentagons will be formed. The corollary of Brianchon's theorem for a circumscribed pentagon is true for all such figures.

## FIVE STRUCTURAL OBSERVATIONS

A five-sided figure will ordinarily be formed by five non-concurrent nonparallel lines in a plane, and therefore the localization of a conic that may function as an evolute is immediately suggested. The general equation of a conic in rectangular cartesian co-ordinates is

$$ax^2 + 2hxy + by^2 + 2gx + 2fy + c = 0$$

If this is divided by a, it becomes

$$x^2 + 2h^1xy + b^1y^2 + 2g^1x + 2f^1y + c^1 = 0$$

where

$$h^1 = \frac{h}{a}, b^1 = \frac{b}{a}, g^1 = \frac{g}{a}, f^1 = \frac{f}{a} \text{ and } c^1 = \frac{c}{a}$$

The five independent constants $h^1$, $b^1$, $g^1$, $f^1$, and $c^1$ show that the character of any conic is determined by five parameters, and that one and only one conic may be drawn through any five points. Instead of five points, however, five tangents to a conic will likewise determine uniquely its locus.

Five observations of dip, along a traverse, will define the positions of five normals to the traces of the stratigraphic surfaces in a profile plane; and these normals may be regarded as tangents to some conic, whose form and position are to be determined. The conic thus located may then be considered as an evolute, from which a set of involutes may be drawn that will represent the traces of the stratigraphic surfaces. To construct graphically this evolute, it is desirable first to locate its points of contact with the five tangents. The traverse, owing to topographic irregularities, will be a curved line in the profile plane, and in actual work will be so plotted. In Figures 3, 4, 7, and 8, however, relief is omitted from the profiles, in order to gain simplicity in the theoretical presentation.

The application of this method to the construction of geologic profiles will now be demonstrated. In figure 3, let AE be the horizontal trace of a vertical profile plane containing the geologic traverse; and let five dips be plotted at the points A, B, C, D, and E. Normals to the dips drawn at these five points will be regarded as tangents to the desired conic. These tangents will intersect as shown, forming the concave pentagon PQRST, outlined by heavy lines. The lines PR and QS are drawn to locate the point $O_1$. The line $TO_1$ is drawn to intersect the extension of the side QR thus locating one point of tangency to the conic. Four similar constructions yield the other points of tangency. The complete procedure is as follows:

Draw PR and QS to locate $O_1$. Draw $TO_1$ to intersect the side QR.

Draw QS and RT to locate $O_2$. Draw $PO_2$ to intersect the side RS.
Draw RT and PS to locate $O_3$. Draw $QO_3$ to intersect the side ST.
Draw PS and QT to locate $O_4$. Draw $RO_4$ to intersect the side TP.
Draw QT and PR to locate $O_5$. Draw $SO_5$ to intersect the side PQ.

Through the five points of tangency, a curve may ordinarily be sketched, which from its mode of construction approximates closely to a conic, and may be used as an evolute. In order not to obscure the constructional lines of Figure 3, this evolute, together with its involutes, are separately shown in Figure 4. The method of constructing involutes from an evolute has already been described in two earlier publications (Mertie, 1940, p. 1107–1134; Mertie, 1947, p. 779–802) and need not be repeated.

The demonstration illustrated by Figures 3 and 4 shows five dips that increase from A to E; but within this interval they may increase at first, then decrease, and again increase, or vice versa. Reversals of dip may also occur within the traversed interval. These conditions, singly and in combination, give rise to the formation of many curiously shaped pentagons. Examples of such figures, flattened to represent the shapes in which they commonly occur, are shown in Figure 5. In applying this method it is necessary first to identify and letter the five-sided figure formed by the intersecting tangents, and thereafter to follow some definite scheme of construction, such as the one tabulated above.

## FOUR STRUCTURAL OBSERVATIONS

The general conic has been shown to depend upon five independent parameters; but for a parabola, the c-derivative of the discriminant vanishes, that is,

$$h^2 - ab = 0$$

This relationship makes it possible to eliminate one of the five parameters of the general conic, so that a parabola may be constructed from four of its tangents, which as before, are normals to the traces of the stratigraphic surfaces.

The construction of a parabolic evolute likewise depends upon the corollary of Brianchon's theorem that applies to a circumscribed pentagon; but four of the circumscribed sides are finite tangents to the parabola, whereas the fifth side is considered to be an ideal tangent at infinity. A parabola, circumscribed by four tangents, is shown in Figure 6. The four finite tangents intersect to form the points B, C, and D. The fifth tangent, AE, lies at infinity. Therefore all lines extended to the ideal point A are drawn parallel to BA, and all lines extended to the ideal point E are drawn parallel to DE. With this convention, the method becomes exactly like that used in constructing the general conic from five tangents.

The application of this method is shown in Figure 7. Let AD be the horizontal trace of the profile plane containing a geologic traverse; and let four dips be plotted at the points A, B, C, and D. Normals to the dips drawn at these four points are also tangents to the desired parabola, and will intersect as shown, locating the points Q, R, and S. The points P and T are considered to be the intersections of tangents at infinity, so that PT is the tangent at infinity. The lines PR and QS are now drawn, the former being parallel to BE; and through their intersection, $O_1$, a line is drawn from T (that is, parallel to ST), to intersect the line QR. This intersection

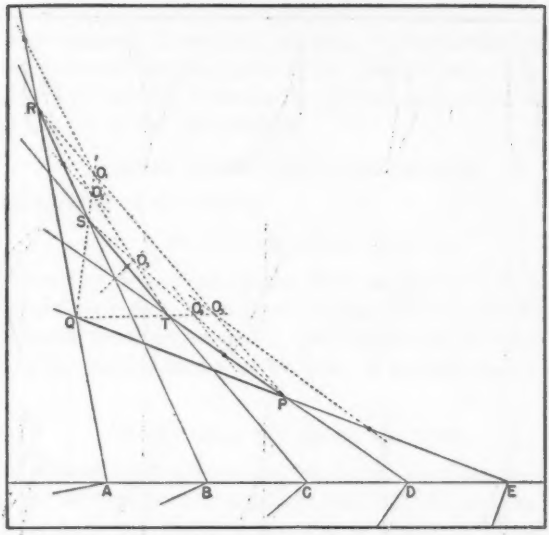


FIGURE 3.—*Concave pentagon produced by five normals to five dips*

Shows construction of points of tangency to inscribed conic

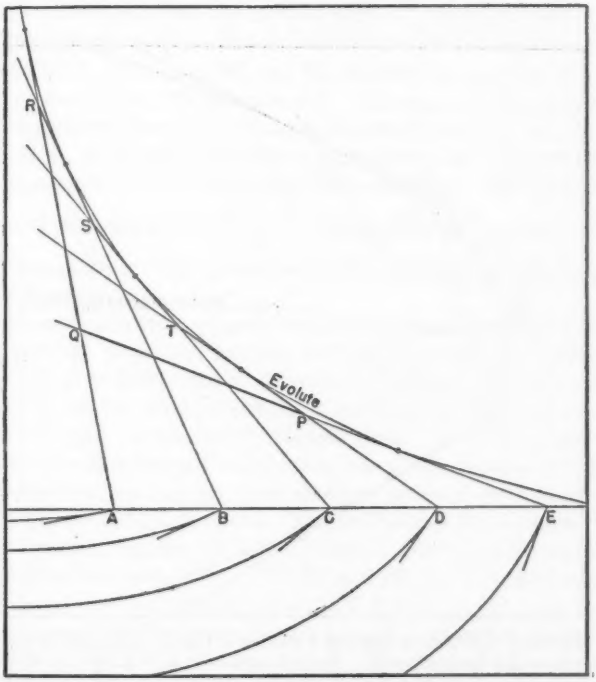


FIGURE 4.—*Evolute and involutes constructed from five normals to five dips*

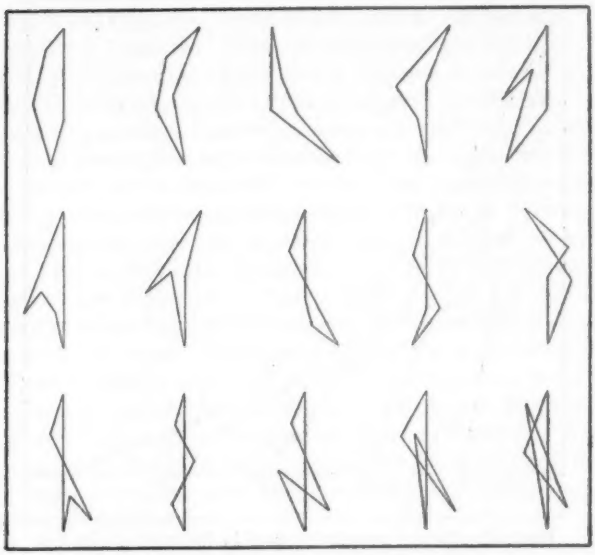

FIGURE 5.—*Convex, concave, proper, and improper pentagons produced by five normals to five dir*

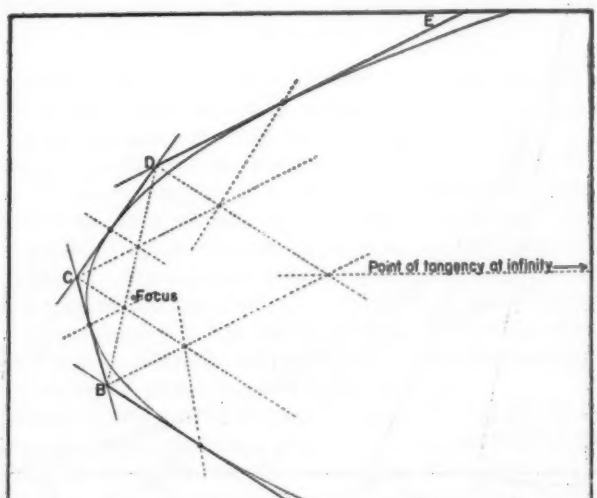


FIGURE 6.—*Pentagon, of four finite lines and a fifth line at infinity, circumscribed about a parabola*

is one point of tangency to the required parabola. Three similar constructions yield the other finite points of tangency; and a fifth construction, connecting R with $O_5$, the intersection of QT and SP, indicates the point of tangency at infinity. The line $RO_5$ is parallel to the axis of the parabola.

## THREE STRUCTURAL OBSERVATIONS

The general equation of the circle is

$$x^2 + y^2 + 2gx + 2fy + c = 0$$

The circle is therefore determined by the three parameters g, f, and c. The construction of a circular evolute from three tangents consists merely in bisecting the two angles between the three tangents. The intersection of these bisectors is the center of the circle, which is used as an evolute. A typical construction is shown in Figure 8.

## ADDITIONAL POINTS ON EVOLUTE

Ordinarily five tangents to a conic, and the five points of tangency, will suffice for sketching the required arc of an evolute. It may happen, however, that two of the points are so far separated that it is desirable to locate one or more intermediate points. This may readily be done by means of Pascal's theorem, of which Brianchon's theorem is an immediate consequence.

Pascal's theorem states that the opposite edges of a hexagon inscribed in a conic meet in three collinear points. Figure 9 shows an ellipse in which is inscribed the hexagon ABCDEF. The sides BC and EF intersect in the point R; AB and DE intersect in Q; and CD and FA intersect in P. The points P, Q, and R are collinear, as required by Pascal's theorem. It should be noted, however, that through the six vertices, A, B, C, D, E, and F, 59 other hexagons may also be drawn; and that for each of these constructions a different line PQR will result. The general rule for the construction of polygons is that $\frac{(n-1)!}{2}$ constructions are possible, where n is the number of given points. This generalization is applicable both to Pascal's and to Brianchon's theorems.

Additional points on a conic, after five are known, are obtained by working Pascal's theorem in reverse. Thus in Figure 10, let A, B, C, D, and E be five points on a hyperbolic evolute, as determined by Brianchon's theorem. The points B and C are considered to be too far apart for sketching satisfactorily the evolute; and it is desired to find some point on the evolute between them. A line ES is assumed in such a position that it will intersect the hyperbola somewhere between B and C; but because the hyperbola has not yet been sketched between these points, the exact position of S remains to be determined. The figure ABCDES, however, is obviously an improper hexagon, because the side BC intersects the sides ES and SA. Under this interpretation, the lines AB and DE, BC and ES, and CD and SA, are opposite sides of the hexagon. Now extend AB and DE to meet in the point R. The side BC and the assumed side ES intersect in the point P. Therefore the line PR, corresponding to PR in Figure 9, is now determined. Then extend CD to meet PR in the

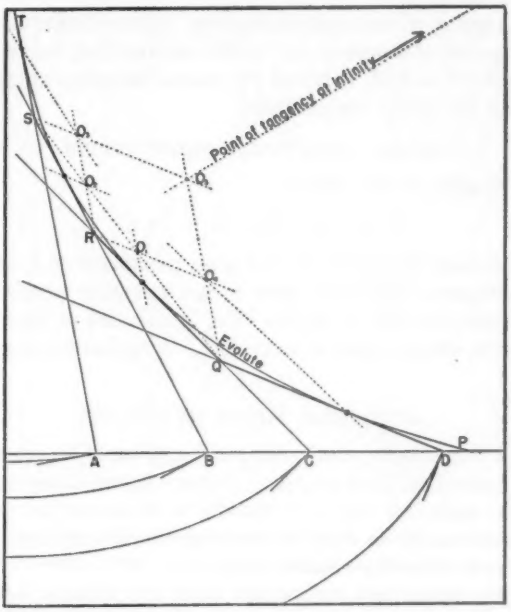


FIGURE 7.—*Parabolic evolute and involutes, constructed from four normals to four dips*

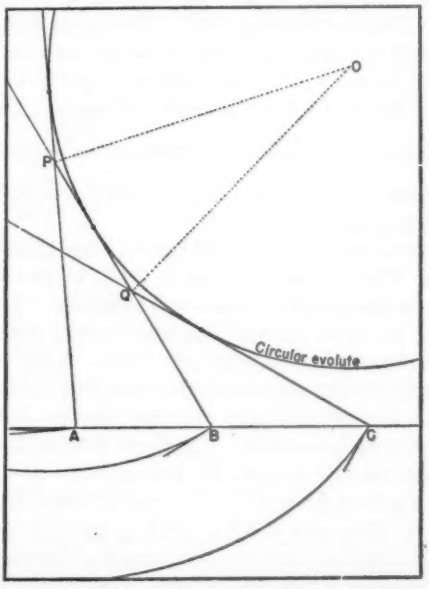


FIGURE 8.—*Circular evolute and involutes, constructed from three normals to three dips*

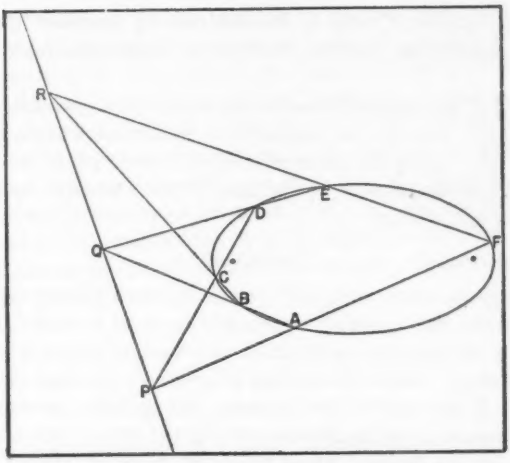


FIGURE 9.—*Convex hexagon inscribed in an ellipse*

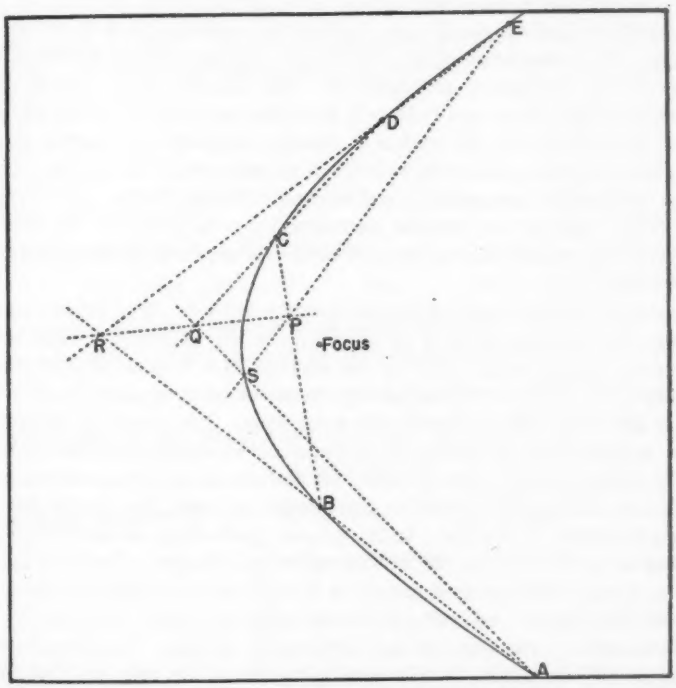


FIGURE 10.—*Application of Pascal's theorem to the determination of additional points on a conic evolute*

point Q. Finally connect Q with A, thus intersecting the point S, which was to be determined. Similarly any number of additional points on a conic evolute may be located.

If four instead of five original observations are given, a parabolic evolute may be drawn (Figs. 6, 7). On such an evolute, four finite points and one infinite (ideal) point were located. Using the convention heretofore described for ideal points and lines at infinity, it is also possible, by the use of Pascal's theorem, to locate additional points on a parabolic evolute.

## ANALYSIS

The methods given above are purely geometric, but algebraic expressions for the evolute or involutes may possibly be required in order to study the curvature of folds, to compare one structure analytically with another, or possibly for the purpose of evaluation of error. Involutes are easy to construct, but their equations are difficult to obtain. If one started, for example, with a single parabolic involute, the equation of its evolute could be derived and charted, after which all the other involutes could readily be drawn. These involutes, however, though parallel to the parabola, are not likewise parabolas, but instead are a family of higher plane curves. To obtain the equations of such families of curves, it is necessary to start with an assumed or determined equation of the evolute, and therefrom to formulate a differential equation, from the quadrature of which the equations of the involutes may be obtained. The formulation of the differential equation may be laborious, and its quadrature may be difficult or impossible. The equation of an evolute, however, represents a single curve, and ordinarily is neither too difficult nor too laborious to derive. An equation for the evolute is therefore considered in itself to be a satisfactory algebraic record of a parallel fold. A general method for obtaining the equations of evolutes has been given in earlier papers (Mertie, 1940, p. 1107–1134; 1947, p. 779–802). Another and simpler method will now be presented for deriving the equation of any conic that is to function as an evolute, from the equations of five of its tangents.

A choice of co-ordinates is the first matter to be decided. If caretsian co-ordinates are used, the horizontal trace of the profile plane will naturally be taken as the X axis; but no natural choice exists for the selection of a Y axis. The length of this horizontal trace, however, can and always should be taken as unity, in order to compare one structure with another on the same scale. The system of trilinear co-ordinates, so much used in charting the physical and chemical properties of rocks and minerals, utilizes the principle of a unit scale and also yields homogeneous equations. Such co-ordinates may be referred to any triangle, or trigon, but the simplest system utilizes an equilateral triangle. In the present application, the horizontal trace of the traverse is taken as the base of this equilateral triangle. The co-ordinate distances $\alpha$, $\beta$, and $\gamma$ are measured normal to the sides of the triangle of reference, and are positive or negative according to whether they are drawn from the sides of the triangle toward or away from the opposite angular vertices. Negative co-ordinates will appear only for points that lie outside the triangle of reference; but not more than two co-ordinates may be negative.

The trilinear equations of the five tangents to the conic must first be written. In Figure 11, the line AC represents the horizontal trace of a geologic traverse, adjusted to a unit scale, so that AC = 1; and the equilateral triangle ABC is the triangle of reference. Let D be any point on the line AC, such that AD = s and DC = 1 − s; also let the dip, $\delta$, and the normal to the dip, be constructed at D. As distances along the sides of the trigon are proportional to the trilinear co-ordinates, the co-ordinates $(\alpha_1, \beta_1, \gamma_1)$ of D may be written as (1 − s, 0, s). Now lay off the angles $\theta_1$, $\theta_2$, and $\theta_3$, measuring these from the normal anti-clockwise to directed lines drawn parallel to the sides of the trigon, that is, to the segments DC′, DA, and DB′. The sines of these angles are the direction sines of the normal. The equation of the normal may then be written as follows:

$$\begin{vmatrix} \alpha & \beta & \gamma \\ \alpha_1 & \beta_1 & \gamma_1 \\ \sin\theta_1 & \sin\theta_2 & \sin\theta_3 \end{vmatrix} = 0 \qquad (1)$$

Two of the normals will pass through A and C, the end points of the traverse. Figure 12 shows the construction at the point A, where s = 0. The construction at C, where s = 1, is similar to that at A.

The co-ordinates of D, in Figure 11, are (.4, 0, .6), and of A, in Figure 12, are (1, 0, 0). The dip, $\delta$, in both figures is 50°, from which it readily follows that $\theta_1 = 160°$, $\theta_2 = 40°$, and $\theta_3 = 280°$. The resulting direction sines are respectively .342, .643, and −.985. Therefore the equations of the normals in Figures 11 and 12 are as follows:

$$\begin{vmatrix} \alpha & \beta & \gamma \\ .4 & 0 & .6 \\ .342 & .643 & -.985 \end{vmatrix} = 0 \qquad \text{and} \qquad \begin{vmatrix} \alpha & \beta & \gamma \\ 1 & 0 & 0 \\ .342 & .643 & -.985 \end{vmatrix} = 0$$

These two determinants yield the equations

$$.386\alpha - .599\beta - .257\gamma = 0 \qquad \text{and} \qquad .985\beta + .643\gamma = 0$$

The equations of the other normals are similarly derived.

The general equation of a conic in trilinear co-ordinates is

$$\phi(\alpha, \beta, \gamma) = a\alpha^2 + b\beta^2 + c\gamma^2 + 2f\beta\gamma + 2g\gamma\alpha + 2h\alpha\beta = 0 \qquad (2)$$

The general equation of a straight line in trilinear co-ordinates is

$$l\alpha + m\beta + n\gamma = 0 \qquad (3)$$

Differentiating equation (2) with regard to $\alpha$, $\beta$, and $\gamma$, we have

$$\frac{\partial\phi}{\partial\alpha} = 2a\alpha + 2h\beta + 2g\gamma$$

$$\frac{\partial\phi}{\partial\beta} = 2h\alpha + 2b\beta + 2f\gamma$$

$$\frac{\partial\phi}{\partial\gamma} = 2g\alpha + 2f\beta + 2c\gamma$$

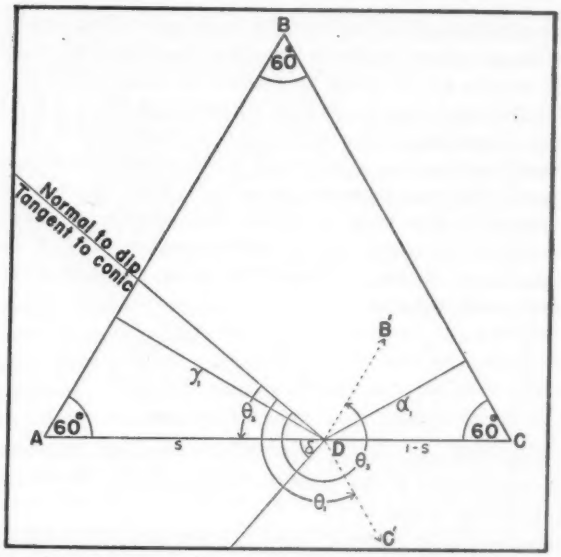


FIGURE 11.—*Equilateral triangle of reference for trilinear coordinates, with application to midpoint of traverse*

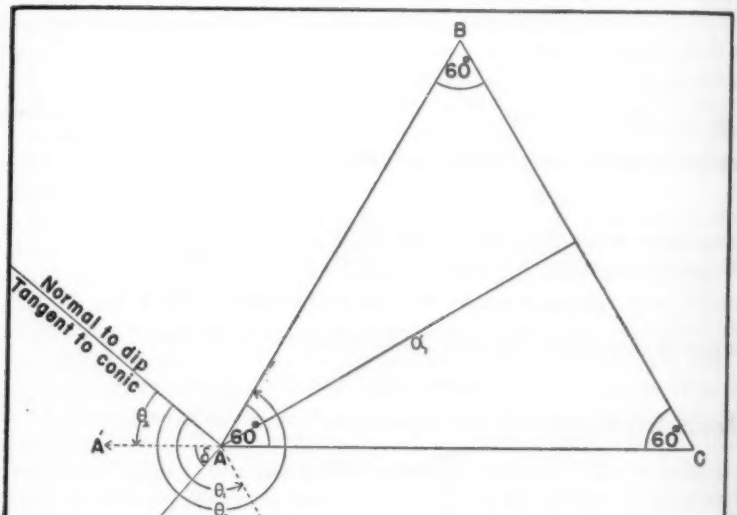


FIGURE 12.—*Equilateral triangle of reference for trilinear coordinates, with application to end point of traverse*

The general equation of a tangent to any curve, at $(\alpha_0, \beta_0, \gamma_0)$ is

$$\left(\frac{\partial\phi}{\partial\alpha}\right)_0 \alpha + \left(\frac{\partial\phi}{\partial\beta}\right)_0 \beta + \left(\frac{\partial\phi}{\partial\gamma}\right)_0 \gamma = 0$$

Therefore

$$(a\alpha_0 + h\beta_0 + g\gamma_0)\,\alpha + (h\alpha_0 + b\beta_0 + f\gamma_0)\,\beta + (g\alpha_0 + f\beta_0 + c\gamma_0)\,\gamma = 0 \qquad (4)$$

Equation (4) has the form of (3), the equation of a straight line, so that by equating coefficients, the following equations are obtained:

$$a\alpha_0 + h\beta_0 + g\gamma_0 = l \qquad (5)$$

$$h\alpha_0 + b\beta_0 + f\gamma_0 = m \qquad (6)$$

$$g\alpha_0 + f\beta_0 + c\gamma_0 = n \qquad (7)$$

But the equation of a straight line through the point $(\alpha_0, \beta_0, \gamma_0)$ is

$$l\alpha_0 + m\beta_0 + n\gamma_0 = 0 \qquad (8)$$

Eliminating $\alpha_0$, $\beta_0$, and $\gamma_0$ from equations (5), (6), (7), and (8), we have

$$\begin{vmatrix} a & h & g & l \\ h & b & f & m \\ g & f & c & n \\ l & m & n & o \end{vmatrix} = 0 \qquad (9)$$

which may be written

$$(bc - f^2)l^2 + (ca - g^2)m^2 + (ab - h^2)n^2 + 2(gh - af)mn + 2(hf - bg)nl + 2(fg - ch)lm = 0 \qquad (10)$$

The five equations of the tangents, obtained from equation (1), have the form of equation (3); and the coefficients of $\alpha$, $\beta$, and $\gamma$, in these equations, are $l_n$, $m_n$, and $n_n$. Therefore five additional equations having the form of (10) may be written, wherein $l_1$, $l_2$, $l_3$, $l_4$, and $l_5$ are substituted for l, and similar coefficients are substituted for m and n. From these five equations and from (10), the coefficients of $l^2$, $m^2$, $n^2$ mn, nl, and lm may be eliminated to produce the following determinant:

$$\psi(l, m, n) = \begin{vmatrix} l^2 & m^2 & n^2 & mn & nl & lm \\ l_1^2 & m_1^2 & n_1^2 & m_1n_1 & n_1l_1 & l_1m_1 \\ l_2^2 & m_2^2 & n_2^2 & m_2n_2 & n_2l_2 & l_2m_2 \\ l_3^2 & m_3^2 & n_3^2 & m_3n_3 & n_3l_3 & l_3m_3 \\ l_4^2 & m_4^2 & n_4^2 & m_4n_4 & n_4l_4 & l_4m_4 \\ l_5^2 & m_5^2 & n_5^2 & m_5n_5 & n_5l_5 & l_5m_5 \end{vmatrix} = 0 \qquad (11)$$

This determinant may of course be expanded by the use of minors, but several shorter methods are known, of which one of the best is given by Muir (1933, p. 63–66). The Chiò method of pivotal condensation, said by Aitken (1942, p. 45–48) to have first been used by Gauss, is also useful.

A straight line may be defined by three normals drawn from the vertices of a tri-

angle of reference, which in this application is an equilateral triangle. The lengths of these lines, shown in Figure 13, are the tangential co-ordinates (l, m, n) of the straight line. By regarding l, m, and n as variables, any straight line may be thus represented. If these co-ordinates are related by some equation in l, m, and n, this

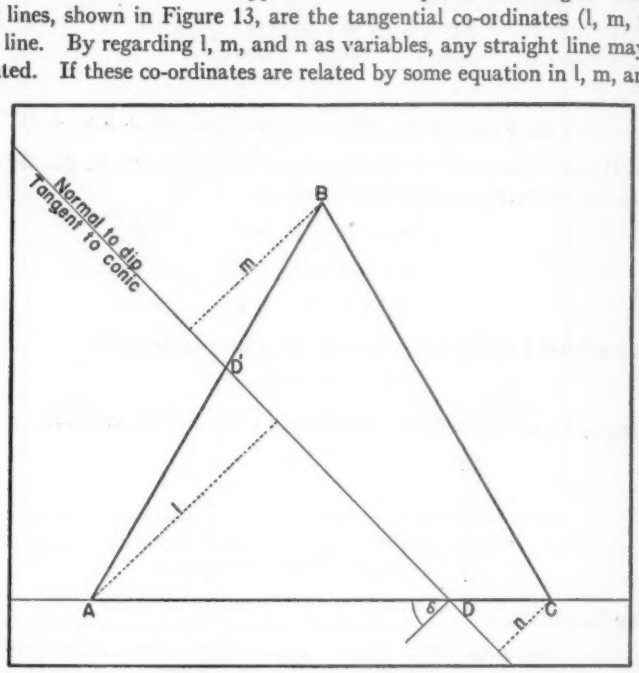


FIGURE 13.—*Equilateral triangle of reference for tangential coordinates, with application to a straight line*

relationship is known as a tangential equation. The function $\psi$ is the tangential equation of the family of straight lines that are tangent to a conic. This function is therefore not the required equation of the conic in trilinear co-ordinates, but it may readily be converted thereto. Thus, when the determinant equation (11) is expanded and re-assembled, it will have the form of

$$\psi(l, m, n) = Al^2 + Bm^2 + Cn^2 + 2Fmn + 2Gnl + 2Hlm = 0 \qquad (12)$$

This may be converted to the equation of the conic in trilinear co-ordinates by means of the following determinant, which is the analogue of (9).

$$\omega(\alpha, \beta, \gamma) = \begin{vmatrix} A & H & G & \alpha \\ H & B & F & \beta \\ G & F & C & \gamma \\ \alpha & \beta & \gamma & o \end{vmatrix} = 0 \qquad (13)$$

Expanded and re-assembled, this will have the form of

$$\omega(\alpha, \beta, \gamma) = L\alpha^2 + M\beta^2 + N\gamma^2 + 2P\beta\gamma + 2Q\gamma\alpha + 2R\alpha\beta = 0 \qquad (14)$$

The objective stated for this analysis was the derivation of an algebraic expression that would record the parallel folding indicated by five normals to five dips, and would enable one fold to be compared with another. Equation (14), in trilinear co-ordinates, accomplishes this objective, but so does equation (12), written in tangential co-ordinates. The choice between the two equations depends upon the use to which they are to be put. As records, and also for comparative purposes, equation (12) is as good as (14), and saves additional algebraic work. If the conic is to be charted from its equation, however, equation (14) is to be preferred.

Equation (14) will commonly represent either a hyperbola or an ellipse, as the chance is small that the eccentricity will be exactly unity. The true character of the conic, however, may be determined by examining the $\omega$ function of the lengths of the sides of the trigon, that is, $\omega(1, 1, 1)$. If this function is positive, the conic is a hyperbola; if it is negative, the conic is an ellipse; and if the function vanishes, the conic is a parabola. If $\omega(O, N, -M) = \omega(-N, O, L) = \omega(M, -L, O)$, or otherwise stated, if $M + N - 2P = L + N - 2Q = L + M - 2R$, the conic is a circle.

It may be desirable to transform some of these trilinear equations to rectangular cartesian equations. To do this, an origin is selected whose trilinear co-ordinates are $p_1$, $p_2$, and $p_3$. If the angles between the X axis and the three sides of the trigon are $\zeta_1$, $\zeta_2$, and $\zeta_3$, the formulae for the transformation are as follows:

$$\alpha = x \cos \zeta_1 + y \sin \zeta_1 - p_1$$

$$\beta = x \cos \zeta_2 + y \sin \zeta_2 - p_2$$

$$\gamma = x \cos \zeta_3 + y \sin \zeta_3 - p_3$$

## COMPARISON OF METHODS

The Hewett method of concentric arcs, or involutes from a point, was designed primarily for the measurement of stratigraphic thickness, in suitable vertical plane profiles normal to the strike of the rocks. Restricted to sections where the strike is constant, the method is inexact, because the traces of parallel stratigraphic surfaces are not ordinarily concentric arcs. By the method of evolute and involutes, likewise restricted to suitable plane profiles normal to the strike, the curvature, instead of being inferred or assumed, is actually deduced from observations of dip; and instead of utilizing these dips in pairs, any number of them are used simultaneously, the results improving with the number of observations.

Busk (1929) attempted to use the method of concentric arcs for the construction of profiles; but the method is not suited to this purpose, because the observations must be used in pairs with the result that discontinuities appear in the traces of the parallel stratigraphic surfaces. The method of evolute and involutes is particularly suited to the construction of profiles, because the parallel curves or involutes that represent the traces of the beds are continuous throughout any specified interval.

If the Hewett method is used outside of a vertical plane profile normal to the strike of the rocks, or if aberrant strikes are projected into this plane, the method becomes an empirical one for the calculation of stratigraphic thickness. The same is true for the method of evolute and involutes. For this reason, numerous other methods of calculating stratigraphic thickness have been devised to apply where traverses must

be made oblique to the strike of the rocks. In all such methods so far proposed, observations must be used in pairs; and on this account, the Hewett method of concentric arcs has important indirect applications. The writer (Mertie, 1940, p. 1107–1134; 1947, p. 779–802) has already presented two empirical methods wherein the concept of concentric arcs is utilized.

The general method of evolute and involutes, described in the two earlier publications, appears to be the most universally applicable method for the mechanical construction of plane profiles of parallel folds, where more than two observations of dip are available. The evolutes are higher plane curves that can be represented approximately by the general formula $y = ax^n$. The evolute and involutes are simply drawn by geometrical construction; but in applying this method, it is desirable first to prepare a graph, showing the mode of variation of the dips within the traversed interval. By the use of this variation or interpolation chart, it is possible to draw as many tangents to the desired evolute as may be needed for its accurate delineation. An evolute constructed in this manner yields involutes that are a true representation of the traces of the bedding planes. The method therefore uses to the best possible advantage all of the available data; and although the preparation of a variation chart requires additional work, the results obtained are unexcelled by any other method.

The special method of evolute and involutes, as proposed in this paper, is a modification of the general method, where five, four, or three suitable observations are used. Evolutes are prepared directly from the available data without the use of interpolation diagrams. These conic evolutes have to be sketched from determined points that may be farther apart than is desirable; but if they are too far apart to be useful, additional points may be determined. Yet it still is true that these points are arbitrarily adjusted to produce a conic, whereas in the general method the evolute is a higher plane curve that fits exactly the field data. Therefore the conic evolute is somewhat less accurate, and the derived involutes are less exact. The special method, however, is quicker; and under favorable conditions is sufficiently accurate for most stratigraphic purposes. The conic based on five tangents is of course a better evolute than the one based on four tangents, and the circular evolute is the least desirable. The conditions that favor these shortened methods are regular and short spacing of the structural observations along the traverse, and a sequence of dips that increase or decrease continuously throughout the traversed interval.

These special methods, though designed to fit five, four, or three observations of dip, are as universally applicable as the general method of evolute and involutes. Thus if ten observations are available, two conic evolutes may be constructed; and it will be found that these curves will join one another with little adjustment. If a guide to this adjustment is needed, two tangents from the right suite of five and three tangents from the left suite of five may be selected to construct an intermediate evolute that will join the other two. Similarly, for any number of observations, combinations of different conic evolutes may be utilized to produce a composite evolute. Several conics, however, that are thus adjusted and joined together, will no longer be a conic, but instead will be some higher plane curve.

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Abstract.
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VOL. 59, PP. 787-794, 4 PLS. AUGUST 1948

# EFFECT OF ATOMIC BOMBING ON BUILDING MATERIALS AT HIROSHIMA, JAPAN

BY ALLEN C. TESTER

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## ABSTRACT

Ten months after the atomic bombing of Hiroshima, Japan, the area was examined for the effect on building materials and structures, and to compare the destruction with that of cities attacked with incendiary bombs. The characteristics of the bomb explosion are explained as there has been some misunderstanding of its effects.

The nature of the terrain permitted a spread of the force of the explosion but many variations in effects were seen. The effect of the heat wave is discounted as no evidence indicates high temperature beyond a small fraction of a second, though radiant heat was effective in unusual ways. The pressure wave produced an effective force that was vertical in the ground zero area and spread to nearly horizontal with distance.

Two significant conditions produced by the pressure wave were observed. Air was compressed and drove solid particles with force that abraded stone and steel. Near ground zero concussion developed compression fractures in stone and heavily reinforced concrete. It is believed the pressure developed was above the crushing strength of an average structural granite.

## INTRODUCTION

On August 6, 1945, a new weapon was used in warfare at Hiroshima, Japan. Two reports by War Department survey groups have been published and contain much significant data. The published conclusions on the effect on building materials do

not include all the possibilities, and it is apparent that some observations are subject to different interpretations.

## THE CITY

The general location of Hiroshima near the western end of the island of Honshu is well known. The major portion of the city is situated on the Ota River delta which has been built into Hiroshima Bay on the northwest side of the great Inland Sea (Pl. 1, fig. 1) The six distributaries of the Ota River and two small tributaries at the margins spread like the ribs of a fan, and except for Enko River branch on the extreme southeast, all branches cross alluvial deposits without intervening hills or outliers. Near the southeast corner of the delta area and lying between the Enko and Kyobashi River branches there is one small hill of granite about 200 feet high. Farther southeast and between the same distributaries is an area of about 1 square mile of granite hills with a maximum elevation of about 660 feet. Other than by these two groups of hills the heat and shock waves from the bomb explosion were uninterrupted in their propagation throughout the area of the delta. Along the margins of the area, maturely dissected hills of gray and pink granite rise to ridges 600 to 1100 feet above the delta. The delta sides are each approximately 5 miles long. The head of the delta merges with the flood plain of the Ota River bordered by granite hills for many miles northward.

More than 90 per cent of the maximum wartime population of 380,000 lived on the delta plain with about 60 per cent living in the central area of about 4 square miles. Evacuation programs had reduced the population considerably but in the central congested portion there were probably 35,000 persons per square mile at the time of the bombing (Pl. 1, fig. 2).

The buildings and bridges in Hiroshima included a wide variety of types and materials. Some relatively new buildings of heavy steel-reinforced concrete and massive granite facings on sound foundations were built to withstand earthquake shocks. Bridges with strong piers and heavy steel beam trusses carried the principal traffic-ways over the rivers. But the majority of the buildings and other structures were of typical Japanese construction: one- to two-story frame buildings with delicate sliding window and door millwork and much paper screening, and with tile roofing on roof trusses and framing barely strong enough to carry the load. Many small but rambling industrial buildings were of wood and light-weight steel-frame construction. Brick, tile, or stucco veneer is another very common Japanese type and usually the frame is relatively weak. Standards were changed after the disastrous earthquakes of the past 30 years and especially since the Kwanto earthquake of 1923 which caused extensive damage in the Sagami and Tokyo Bay areas. About 50 commercial buildings in Hiroshima had been built since the more rigid standards were adopted to provide earthquake-proof structures, but practically all of the military buildings were old or temporary structures. Smokestacks, high transmission towers of steel or wood were scarcely and rarely damaged in any way, except where struck by falling walls or driven debris (Pl. 2, fig. 1).

It is well to emphasize that there was no ground shock of any significance as the bomb was detonated several hundred feet above the ground. If the bomb had ex-

FIGURE 1. HIGH VERTICAL, TAKEN 7 AUGUST 1945, DAY FOLLOWING BOMBING
Shows Ota River delta plain, two granite hill areas (lower right-hand), and granite ridges bordering delta. Several small fires may be seen; part of haze is due to smoke. Photo U.S.A.A.F.

FIGURE 2. VERTICAL VIEW OVER GROUND ZERO AT '0'
Several days after bombing. Complete destruction within 1 mile radius except for reinforced buildings which stand in definite relief above the barren areas. Photo U.S.A.A.F.

AERIAL PHOTOGRAPHS OF HIROSHIMA AFTER BOMBING

FIGURE 1. GROUND ZERO OF EXPLOSION IN LEFT MIDDLE-GROUND
Blast pressure was vertical in this area. Steel radio tower on left adjacent to post-office building. None of the buildings shown were earthquake- or blast-resistant. Photo U.S.A. Signal Corps.

FIGURE 2. ELECTRIC COMPANY BUILDING, 2100 FEET FROM GROUND ZERO
Walls, roof, and floors all structurally sound but interior burned by secondary fires. Building used for relief work. Photo U.S.A. Signal Corps.

FIGURE 3. VAULT BUILDING, 200 FEET FROM GROUND ZERO
Steel-reinforced concrete walls 30 inches thick designed to withstand fire, demolition HE bombs, and earthquakes. The vertical blast pressure crushed the walls producing compressional fracture pattern. Photo A.C.T.

VIEWS OF DAMAGED BUILDINGS

ploded on the ground or after penetrating the ground, damage of the type caused by penetration high explosive bombs would have occurred on a very large scale although some of the other effects would have been lost.

## THE EXPLOSION

The atomic bomb is simply a small package with an explosive force far beyond previous experience. The expanding burst is the same as the old HE bombs of the first world war with a sudden and violent release of a huge amount of energy in a large but limited area. The unprecedented amount of energy released by the new bomb takes three forms, and all may be important in destroying structures as well as humans. These forms of energy are:

1. Heat (present in other types of explosives) of intensities and depth of distribution by radiation unknown heretofore.
2. Blast or pressure, as in any high explosive demolition bomb but of a magnitude impossible in a single bomb of previous types.
3. Radiation, a new phenomenon (insignificant or unknown in other types of explosions), the effects of which are not yet wholly understood or evaluated.

It is not yet possible to give a complete quantitative report on the magnitude of the forces developed by the chain reaction which produced the mass release of energy at the moment of detonation of the size and type of bomb used at Hiroshima. Enough is obvious to recognize that a new set of criteria must be established to evaluate the effects on all types of structural materials. President Truman in his radio statement following the Hiroshima strike announced that the bomb *had more power than 20,000 tons of T.N.T.*

In a recent report of the United States Strategic Bombing Survey (1946, p. 24), data on the relative energy of nitroglycerine and uranium show that a unit of uranium has 13,400,000 times the energy of the same amount of nitroglycerine. The heat energy of 1 pound of uranium can heat 25 million gallons of water from freezing to boiling temperature.

The Bombing Survey report (1946, p. 25) further states that the fire ball produced by the explosion was "apparently several hundred feet in diameter", "the temperature at the core was millions of degrees centigrade", and "even at its edge, the temperatures was several thousand degrees." According to this report, instantaneous ignition of cotton and woolen fabrics, thatched roofs, and dry-rotted wood was caused by the bomb flash in areas as much as 3500 feet from the ground center of the explosion. Serious third degree burns of humans were reported at distances up to 4500 feet and possibly farther, and in the immediate vicinity of ground zero bodies were charred beyond recognition. It was also reported that roofing tile was blistered at distances as great as 4000 feet; scarring and peeling of granitic blocks occurred almost a miie from ground zero; charred telephone poles (usually of cedar wood) were found over 2 miles from ground zero; humans reported feeling heat as far as 4½ miles away; and some flesh burns were reported nearly 3 miles from the center. All such burns are classed as flash burns as distinct from flame or fire burns.

There are several variables that complicate the intensity of heat relation to distance from zero. Objects less than 1000 feet from zero that were screened by concrete

structures were not damaged by the flash heat Unscreened light-colored or white fabrics were not burned while black cloth ignited. The flatness of the terrain permitted uniform distribution and reflection of the heat waves. Some variation of intensity was caused by buildings or even some very small objects in the path of the heat waves. Shadow effects were found at all distances up to nearly 2 miles from the ground zero. Some inflammable objects were found unburned almost directly below the bomb center.

There is no question that tremendous heat was developed in the center of the explosion for a fraction of a second and that this heat was intense for a ball of more than 1000 feet in diameter. No object was in contact with this ball of fire as it developed several hundred feet above the ground.

Reports were current at the time of the writer's investigation that iron, steel, and other metallic structural materials had been melted, but a search yielded nothing of that type that could be certainly identified as caused by the primary heat wave. Steel pipe in a bridge and a steel radio tower (Pl. 2, fig. 1) within 550 feet of ground zero and showing no evidence of fusion temperature support the conclusion that high temperatures did not exist for more than a fraction of a second at any distance and that for distances over 5 or 6 thousand feet beyond the ground zero, high temperatures did not exist for any period of time. The paint on the steel radio tower was neither blistered nor burned, and the scaling present appeared to be due to normal atmospheric weathering. If fusion temperature existed for any distance, it was for too short an interval of time to accomplish any but the most superficial work. It seems that only highly inflammable objects of low ignition temperature and human flesh were vulnerable to the radiant heat wave which was not of extreme or high temperature.

## THE FIRE

Fires sprang up immediately and almost simultaneously throughout the heavily damaged area within a radius of about 6300 feet from the ground zero. It is my opinion that practically all of these fires were secondary, due to ignition by household charcoal burners, industrial fires, electrical short circuits, broken gas mains, and other miscellaneous causes in the demolished buildings caused by the pressure blast. In the area 600 to 1000 feet from ground zero, the fires in the interiors of the concrete and stone buildings may have been caused by the primary heat wave although most of the secondary causes were also present.

The burned areas of Hiroshima have very much the same appearance as the burned areas of Tokyo, Yokohama, Nagoya, Osaka, Kobe, and Hiratsuka which were attacked with incendiary bombs. At Hiroshima the fires were intense and spread rapidly. High temperature was generated quickly and the fire wind that soon developed closed the gaps to engulf the intervening buildings that had not been fired by either primary or secondary causes. Metal fixtures were melted; bath and roofing tiles were blistered; soft-burned brick was glazed; glassware was fused and made to flow like miniature tongues of obsidian buried in the thin bed of ashes. Granite dimension blocks showed the effects of heat by superficial granular disintegration and spalling of the outer fraction of an inch but in no case were developed large or deep

fractures that weakened the interior of large blocks. Asphalt and even concrete paving and walks were warped and buckled where the flames were concentrated, yet in protected spots no such damage was seen. The same types of heat effects were observed in the other cities that were attacked with only incendiary bombs.

## THE PRESSURE WAVE

The second major form of energy developed by explosion is the pressure wave. No quantitative data are available at this time on the amount of pressure developed by the atomic explosion. It has been calculated that even a single 100-pound bomb detonated at ground level would develop a higher blast pressure for a few thousandths of a second for a very small radius of 18 feet than was attained on the ground by the atomic bomb as used at Hiroshima (Strategic Bombing Survey, 1946, p. 33).

It is believed that the blast pressure of the atomic bomb, like other high explosives, rose very rapidly to a peak, declined slowly, and then fell below atmospheric pressure for about three times the duration of the higher than atmospheric pressure period. The high pressure period of the atomic explosion lasted slightly more than 1 second which is appreciably longer than the positive pressure phase of any other high explosive type. A block buster HE bomb has a high pressure period of only a few thousandths of a second (Manhattan Engineer District, 1946, pt. 1, p. 20).

At Hiroshima the pressure blast pushed buildings and bent weak wood and steel frames to lean away from the explosion center but crushed some structures within a radius of 600 feet of ground zero. The normal effect of other high explosive types is a sharp blow which pierces or cuts materials. The longer duration of the pressure wave of the atomic bomb produces an effect like the crushing force of a vise or press, while the ordinary explosive develops a quick hammer blow.

The negative pressure period which followed the extreme pressures may have, in certain situations, produced a bursting effect on solid or large closed objects. If buildings and other open structures withstood the initial pressure they were not damaged by the longer period of low or negative pressure.

The resultants of the components of forces were from vertical directly below the bomb burst zero to near horizontal at the extremities of the damaged area. At approximately 3 miles from the blast zero, the resultant of the forces was 5 to 8 degrees from the horizontal component. In the intervening areas both vertical and horizontal forces were effective and walls were pushed in as well as roofs crushed from above. This was especially true between 1000 and 2000 feet from the ground zero where some weak buildings were submerged by the pressure wave. The buildings of massive stone or the heavily reinforced concrete walls with a steel frame structure carrying the floor and roof loads withstood the forces very well (Pl. 2, fig. 2). An example of the crushing vertical force was seen in a vault that was about 200 feet from the ground zero (Pl. 2, fig. 3). The small cubical building had concrete walls about 30 inches thick heavily reinforced with steel rods but the force produced typical compressional fractures in the walls on all sides.

The shielding effect of buildings was negligible as building heights had been restricted to 100 feet because of earthquake hazards. The topography of the region offered no protection except in the extreme southeast part of the city, which was a

small and sparsely occupied area. At Nagasaki many hills and ridges separated the populated areas, and due to this protection the area destroyed at Nagasaki was only 1.8 square miles as compared to 4.7 square miles at Hiroshima (Strategic Bombing Survey report, 1946, p. 33).

Reflection or diffraction effects were of minor importance at Hiroshima although numerous small examples were noted. The reflection of the pressure wave by the water surface of the rivers within 1 mile of the blast center caused considerable damage to bridges as they were not built to withstand upward or oblique forces from below. Diffraction of pressure waves may explain the ability of numerous well-built tall smokestacks to withstand the blast. Very few of the stacks were destroyed or damaged in any way to impair their efficiency.

In summary, the pressure effect was disastrous to weakly constructed buildings within a radius of $1\frac{1}{2}$ miles of ground zero, and much damage was done to the residential and single-story industrial buildings of thin shell or light steel-frame construction as far as 3 miles from the center. But the efficiency of the blast pressure was reduced rapidly beyond this zone (Pl. 1, fig. 2; Pl. 3, fig. 1). In Hiroshima the reinforced concrete and heavy steel-frame buildings designed to withstand earthquake shocks were damaged beyond utility *only* within 500 feet of ground zero, and in four cases the buildings were structurally sound within 800 feet of ground zero (Pl. 3, fig. 2).

## RADIATION

Little is known of the effect of radiation on structural materials. The Strategic Bombing Survey report (1946, p. 28) mentions two reactions that might be significant in this study, namely:

1) a damaging penetration from high-frequency radiations, whether neutrons, gamma rays, or other unspecified rays, released in the chain reaction of the bomb.
2) an induced radioactivity in the bombed area caused by interaction of neutrons with matter penetrated.

Examination of the soil at Hiroshima shows that vegetable and animal life appeared to be normal soon after the bombing. By June 1946 the trees and shrubs were in full leaf in all areas observed and only those trees that had been severely damaged, killed, or broken by falling or driven debris were barren. What the effect would have been had the bomb exploded on the ground is a conjecture. Radioactivity was found in soils 10,000 feet from ground zero several weeks after the explosion. It is possible that this was from deposits of primary fission products. It is known that the sands from the New Mexico site of the first atomic bomb explosion are still radioactive. Specimens of rocks collected by the writer from buildings at Hiroshima were subjected to Geiger counter and electroscope tests in September 1946 and found to be negative.

## EFFECTS ON STRUCTURAL MATERIALS

Some monumental and structural blocks of granite and syenite show a surface with pronounced pitting and roughening on the sides facing the ground zero. The sur-

FIGURE 1. HIGH OBLIQUE LOOKING WESTWARD
Circle diameter about 190 feet, ground zero at center. Military target in right hand area. Buildings 5, 10, 14, and 15 structurally sound; 3, 6, and 9 badly damaged and worthless; all had secondary fires. Photo U.S.A.A.F.

FIGURE 2. VERTICAL DIRECTLY OVER GROUND ZERO
Buildings of figure 1 can be identified. Roof damage indicated by irregular dark areas; regular dark areas are airways or open shafts. Photo U.S.A.A.F.

AERIAL VIEWS NEAR GROUND ZERO

FIGURE 1. CORNER OF MONUMENT BASE
About 1300 feet northwest of ground zero. Light side at right faced zero; abrasion by air and dust blast extends around corner and concentrates in channels; dark area at top unabraded. Black igneous rock at base. Photo A.C.T.

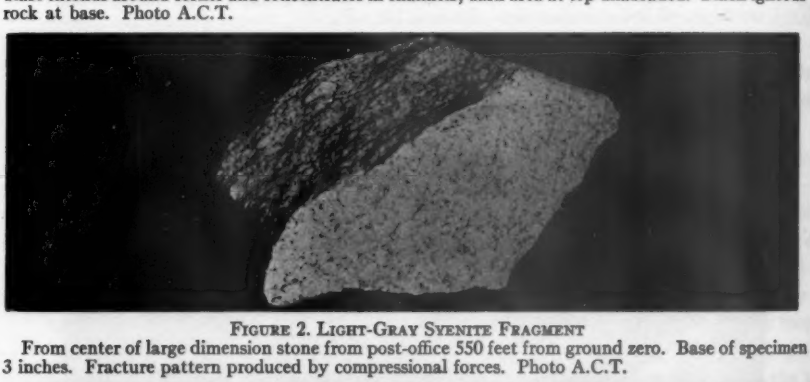

FIGURE 2. LIGHT-GRAY SYENITE FRAGMENT
From center of large dimension stone from post-office 550 feet from ground zero. Base of specimen 3 inches. Fracture pattern produced by compressional forces. Photo A.C.T.

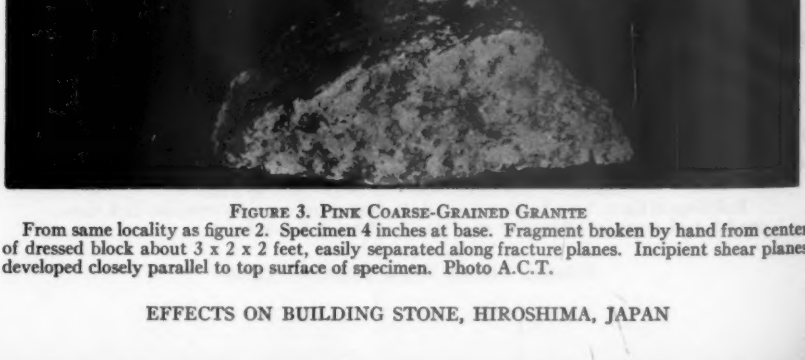

FIGURE 3. PINK COARSE-GRAINED GRANITE
From same locality as figure 2. Specimen 4 inches at base. Fragment broken by hand from center of dressed block about 3 x 2 x 2 feet, easily separated along fracture planes. Incipient shear planes developed closely parallel to top surface of specimen. Photo A.C.T.

faces might also be described as typical of irregular granular disintegration with depth due to atmospheric weathering in which frost action had been especially effective. This phenomenon has been interpreted by others as due to heat which produced a rapid expansion of the outer mineral grains and chipped the corners of the blocks and sharp edges of the tooled grooves (Manhattan Engineer District, 1946, pt. 1, p. 27). However, within inches of the major pitted and broken surfaces but around the corner in a protected position away from a straight line travelled by radiant heat from the blast center, the pitted condition continues and gradually fades into the original highly polished surface; the block is structurally sound. The writer does not believe that a large temperature differential would exist within 6 to 8 inches, nor is it clear why the markings should show a pattern of distribution characteristic of air flow if they were caused by radiated heat. The condition can be explained better as being due to a process similar to sand blasting, or corrasion by concussion. The principal corraded surface (Pl. 4, fig. 1) faces the ground zero and extends around the corners but with diminishing intensity in the lee of the corner.

Concussion, or the shock produced by the collision of bodies, would be in this case the impact of highly compressed air striking head-on the rock surface in the direct path of the blast. That force alone might be sufficient to crush the surface zone of mineral grains of the rock and to weaken the binding effect of their natural interlocking structure. In addition, there would be a striking force delivered by the particles of dust in the atmosphere prior to the explosion. The weather had been dry and hot in Hiroshima prior to August 6th, and the air undoubtedly contained the normal amount of dust raised from the silty clay soil of the alluvial plain. Such material normally distributed throughout the air of pre-blast density would be concentrated and impelled with strong velocity by the compression of the air following the blast. Additional dust and even larger solid particles probably were picked up as the air wave passed downward and outward from the point of detonation.

How high the air temperature might have been is uncertain, but all the evidence shows it was below fusion of minerals, steel, and ceramics. Disintegration of rocks is accomplished by a relatively low degree of heat when continued for many hours or repeated frequently, but this single heat blast lasted for only a second or less.

Similar conditions were observed on other standing objects. Iron and steel electroliers were slightly burnished on the side facing the blast, the paint or rust being removed and a small degree of polishing accomplished. The concrete of smokestacks was noticeably roughened with some cases of actual pitting on the side facing the blast.

The Imperial Government post office and telegraph building was located less than 500 feet from the ground zero of the explosion. At this point the blast forces were practically vertical. The building being of weak construction was crushed and much of the brick and stucco material pulverized. No evidence was seen of heavy fire at this location. Two types of granite had been used in the building in the foundation course and for window and door trim. Samples of both types have been examined for heat and pressure effects.

One of the specimens is a medium-grained, uniform-textured, light-gray syenite composed of orthoclase, albite, biotite, hornblende, and a very small amount of quartz

(Pl. 4, fig. 2). The specimen came from several inches below the dressed surface and near the center of the block, and all the minerals are fresh, essentially unaltered, and with normal boundaries of crystallization. The rock was quarried in Hiroshima Prefecture and is of the same type as the Nagoya "granite", a fine and sound building stone used commonly throughout Japan.

The second rock specimen from the post office is very similar in appearance to the Missouri granite from St. Francois County of the Ozark Mountains (Pl. 4, fig. 3). The Japanese rock is a coarsely crystalline porphyritic pink granite of orthoclase, quartz, biotite, and a small amount of hornblende. The source of this rock was not determined with certainty but it is believed to have come from Yamaguchi Prefecture of western Honshu.

Several blocks of both types of rock were examined and each showed the same physical effects of the blast. The outer surfaces of the blocks showed no effects of high temperatures and none of the minerals had been fused as reported by other observers at Hiroshima. Large pieces of rock were easily broken from dressed stone of dimensions up to 5 by 3 by 2 feet. The outer portion of the larger blocks could be separated by hand for depths of 8 to 10 inches. Beneath the outer zone of fracture and toward the center, the stones were so weakened that light hammer blows produced nearly complete mass fragmentation of the rock. The interior of the block separated along planes that are similar to shearing fractures due to compressional stresses. Assuming the rocks to have a crushing strength within limits of the averages of tested granites and syenites, it would suggest that the vertical forces developed a pressure in the range between 20,000 and 30,000 pounds per square inch (1400 to 2100 kilograms per square centimeter).

The general conclusion is reached that heat was not a primary cause of structural failure of the granite dimension blocks. Additional information was gained from the steel radio tower at the post-office site. The steel was free of any markings other than a slight burnishing. Some paint had peeled but none was blistered and soft wire showed no heat effect. As the blast force was nearly vertical downward at this point the slender triangular tower was not distorted nor damaged in any way. By the same downward forces the granite dimension stones were subjected to a strong vertical pressure which crumbled the weaker material and produced compression fractures in the stronger granite. The concrete walls of the nearby vault (Pl. 2, fig. 3) show the same fracture pattern. Concussion, rather than high temperature, is believed to be the most significant force responsible for the major strains and disintegration developed in the building materials at Hiroshima.

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# DONNER PASS ZONE OF DEFORMATION, SIERRA NEVADA, CALIFORNIA

BY FRANK S. HUDSON

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## ABSTRACT

The trough extending north from Tahoe, past Donner Lake, California, is bounded on the west by a zone of uplift in which the base of Tertiary andesitic deposits on the eroded surface of the basement complex is 800 to 2400 feet higher than in adjacent parts of the depression. This paper deals with a part of the uplifted zone extending from Donner Pass 10 miles S. 15° E. to Squaw Peak. Here are two areas of maximum uplift caused by folding and faulting of the east limbs of the folds. The faults with greatest throw dip east at high angles,[1] but notable movements occurred on west-dipping and vertical faults, and, on some faults, the dip changes from east to west in passing along the strike. Deformation began with folding of the entire zone. This was then accentuated, with accompanying faulting, to produce the two structural highs. Finally there was major faulting along the east edge of the zone. Deformation is believed to be post-Miocene and preglacial. The evidence from this region gives no support to the idea that the Tertiary deposits of the Sierra Nevada were deformed merely by westward tilting of a rigid block.

## INTRODUCTION

The men who mined the Tertiary gold-bearing gravels of the Sierra Nevada were acutely conscious of the gradients of the channels. They must soon have learned

[1] Additional field work, after presentation of this paper at Cordilleran section meeting on April 11, 1947, showed the fault of greatest offset to be a west-dipping reverse fault throughout its course. The text and map have been changed accordingly.

that there are changes, sometimes abrupt, in both amount and direction of slope, and that the character of the gravels does not change with varying channel gradients, the deposits of the gently inclined reaches being similar to those found in stretches with gradients in excess of 200 feet per mile. Lindgren (1911, p. 44–50) was the first to attach significance to these facts. He stated that the maximum gradients are in those stretches where the streams flowed southwest, that the gradients of the northwest-trending reaches are moderate, and that those of the northeast-flowing courses are reversed. He concluded that, after the cutting and filling of these channels, the Sierra Nevada range was tilted as a rigid block on an axis trending south-southeast–north-northwest. The amount of tilt in the latitude of the Tertiary Yuba River was, according to Lindgren (1911, p. 154), about 65 feet per mile.

Recent work by the writer showed that the gradients of Tertiary Yuba River west of Dutch Flat cannot have resulted solely from tilting of the whole range as a rigid block. The evidence indicates local deformations of the channel but does not show whether from folding, faulting, or both. Most of this region is covered by heavy soil and brush, and consequently the first study of Tertiary deformation was confined to the barren crestal region of the range. The area selected is at the head of Tertiary Yuba River and includes a 10-mile stretch of the crest from Donner Pass southeast to Squaw Peak.

The geologic map (Pl. 1) is based on stereoscopic study of air photos obtained from U. S. Forest Service, supplemented by 2 weeks field study, in the fall of 1946 and spring of 1947. The map is constructed from U.S.F.S. quadrangle maps. Elevations were calculated by parallax methods, and adjusted to data of U.S.F.S., U.S.G.S., and S.P.Ry. points.

## GEOLOGIC FORMATIONS

### GENERAL STATEMENT

In this part of the range a sequence of Tertiary volcanic deposits rests on the eroded surface of the pre-Tertiary basement complex without the intervention of the non-volcanic gravels, sands, and clays of earlier Tertiary age that are a prominent part of the superjacent series on the lower western slopes. The basement rocks are deep-seated intrusives of the Sierra batholith, enclosing masses of metamorphic rock, considered by Lindgren (1897, p. 2) as part of the Juratrias Sailor Canyon formation. The overlying volcanics include tuffs, agglomerates, and flows, largely andesitic, but locally, at the base, rhyolitic. The andesites are believed to be of upper Miocene or lower Pliocene age (Louderback, 1924, p. 19). The rhyolitic deposits are probably lower Miocene.

Evidence of adjacent areas indicates that the volcanic deposits along the crest of the range were originally at least 2000 feet thick, and that they were part of a continuous mass of great extent. The headwater streams of the American and Yuba river systems west of the crest have eroded the volcanic materials completely from considerable areas, but the tributaries to Truckee River on the east slope have exposed the basement rocks only at their headwaters. A continuous volcanic cover of 1500 feet maximum thickness extending 8 miles southeast from Donner Peak and an isolated mass around Squaw Peak remains on the crest.

Glacial deposits and minor amounts of alluvial materials are present along the stream courses.

### BASE OF THE VOLCANIC SERIES

The surface of the basement under the volcanic mantle is rough in detail, but in the numerous places near the contact, where bedding is discernible in the volcanics, it is essentially parallel with the contact. This indicates that the erosion surface on the basement was essentially horizontal. Two localities were found where the earliest of the volcanic materials appear to have filled gentle depressions. One of these, on the west slope of the main ridge 1 mile west of Anderson Peak, appears to have been a northeast-southwest trending trough, half a mile wide and 200 feet deep. The other is on the east side of the crest, from ½ to 2 miles south-southwest from Donner Peak. Here, field studies show that 170 to 200 feet of rhyolite tuff lies between the andesite and basement. The deposits of the first locality are probably also rhyolitic.

## STRUCTURE

### GENERAL STATEMENT

The general form of the contact of the volcanics on the basement is a ridge with a gentle west slope and steeper east slope. Its crest follows closely the main divide of the present range. Lindgren (1897, p. 3–4) believed that this ridge represented the prevolcanic crest of the Sierra Nevada, and that the volcanic materials were erupted from vents along the summit, filling the prevolcanic depressions, including a deep valley which extended northward from Lake Tahoe. He mentioned Mount Lincoln, Tinker Knob, and Squaw Peak as some of the foci of the eruptions. He recognized no post-andesitic faulting in this part of the Sierra Nevada.

Louderback (1924, p. 23–24) discovered post-andesitic faulting along the east flank of the crest, presenting evidence of faults of 1500 feet aggregate downthrow to the east in the upper parts of Squaw and Bear valleys on the southeast flank of Squaw Peak.

The writer believes that the form of the contact of the volcanic deposits on the basement has resulted mostly from deformation in a zone which extends slightly west of north along the crest and east flank of the main range. Within the area studied, this zone extends S. 15° E. from the vicinity of Donner Pass to Squaw Peak, and it is concluded that the movements within it included both folding and faulting. Two loci of maximum deformation have been recognized; (1) the Northern Uplift, near Donner Pass, and (2) the Southern Uplift, culminating 1 mile northwest of Squaw Peak. The two are separated by a sag that passes northeast across the range a short distance south of Tinker Knob. The glacial deposits have not been deformed and none of the steep slopes of the region appear to be the direct effect of faulting. It is concluded, consequently, that the latest movements were preglacial. The first movements which deformed the andesitic deposits were probably no earlier than Pliocene and may have occurred in late Pliocene time.

### SOUTHERN UPLIFT

From its easternmost exposure, 2 miles northeast of Squaw Peak, the base of the volcanics rises 2000 feet to its high point northwest of the peak. There are numerous

faults along this course, but, as shown in structure section C–C′ (Pl. 1), the aggregate of the displacements on them is less than one-quarter of the total rise; the balance is due to the easterly dip of the basal contact.

In the easternmost three-quarters of the section, the base of the volcanics is cut by four east-dipping normal faults of the fault zone discovered by Louderback (1924, p. 23–24) a short distance to the south. The aggregate of their vertical separations is 380 feet; the remaining 800-foot rise of the contact within the bounding faults of the zone is due to the easterly dip of the volcanics on the treads of the blocks.

The faults in the area west of the normal fault zone account for only 150 feet of the 1200-foot total rise of the base of the volcanics between the westernmost normal fault and the culmination. They vary in trend, direction of apparent movement, and direction of dip, and this variability exists along certain of the individual faults. Some of them persist but a few thousand feet along strike and penetrate upward into the volcanics only a short distance. One fault of this sort, the first one west of the normal fault zone in section C–C′, appears to terminate exactly on that contact at the line of section, but 1 mile to the north it cuts the volcanics. With minor exceptions these faults are reverse faults, dipping west at angles between 70° and vertical. They are believed to indicate the action of compressive forces.

The base of the volcanics west of the normal fault zone has the shape of an anticline with crest at 8500 feet above sea level. Its gentle west limb terminates against a northeast-trending fault. Its steeper east flank is composed of two subsidiary features of anticlinal aspect. There is ample evidence of essential parallelism between bedding within the andesitic series and the base of those deposits. Accordingly it is concluded that the form of the base of the volcanic deposits is the result of post-andesitic deformation.

The steeper dips are adjacent to faults, but neither the amount nor direction of movement on the faults suggests that the dips were caused by drag. It is concluded that deformation was by folding, accompanied in its late stage by minor reverse faulting. The basement complex must have been deformed concurrently. Part of its change of shape was effected by movement on the faults that cut the base of the volcanics, but these will not account for all the deformation. The granite has not suffered widespread crushing, so presumably the remainder of its deformation was caused by small movements on faults that either did not cut the base of the volcanics or caused no noticeable displacement.

Three members of the reverse fault group terminate against the western fault of the normal fault zone at points 2 miles north of Squaw Peak. As the normal fault is not offset or bent at these points, it is believed to be the younger.

### NORTHERN UPLIFT

The dominant feature is a reverse fault which enters the area from the north and extends S. 20° E. to the vicinity of Anderson Peak, where its course swings to slightly west of south. The west side is upthrown; dip is 65°–87° W. Maximum vertical separation of about 1500 feet is in the latitude of Donner Pass.

Post-andesitic deformation west of the major fault is proved by the essential parallelism of bedding to base of the volcanics, regardless of the amount or direction of dip.

Faulting accounts for part of the deformation, but there must also have been folding. The steep east dips along the major fault northeast of Anderson Peak (structure section B–B′) might be ascribed to drag, but this could not have caused the steep west dips north of Donner Peak (structure section A–A′) nor the anticline which extends northward from Tinker Knob through Anderson Peak.

Along this anticline the base of the volcanics attains a high point at 7900 feet above sea level in the area of basement outcrop adjacent to the major fault, 2 miles south-southeast of Donner Peak. This is a subsidiary high on the Northern Uplift, separated from the main culmination to the north by a sag, within which is an east-west zone of minor normal faults, which dip steeply southward. These are clean-cut breaks directly across structural features produced by the folding and must have been formed after the folding. They cannot be younger than the major fault because it shows no offsets or bends where they terminate against it. Accordingly it is concluded that the major fault is younger than the folding.

### ORIGIN OF THE UPLIFTS

The two uplifts stand in echelon, both as to the faults which bound them on the east and as to the structural features west of these faults. They are units with similar structural histories within a zone of deformation which extends, probably for many miles, both north and south of the area of study. Their situation can hardly be fortuitous. A fundamental structural feature must have extended continuously along the zone.

There is no suggestion of any relation between the position of this zone and contacts between basement rocks of different physical properties, nor is there evidence of a continuous fault along the zone. Accordingly it is concluded that the fundamental feature, which controlled the position of the later features, was an anticlinal fold, continuous in extent along the zone of deformation.

The variability of direction of dip on the major fault of the Northern Uplift suggests that the block which this fault bounds on the east has been upthrust vertically. The normal faulting of the Southern Uplift may be of the same type. In discussing fault movements of this kind Willis (1946, p. 1882) showed how the upward drive of a wedge, bounded by outward-sloping normal faults, could induce local compression and folding. Such action appears to be mechanically possible and may be, as Willis proposed (1946, p. 1880), the explanation of the basin ranges cited by him. But this idea does not seem applicable to the Donner zone where folding preceded faulting.

### TILTING OF THE SIERRA NEVADA

The principal conclusion of the present study is that tilting of the range as a rigid block cannot explain the structure of the superjacent Tertiary deposits. The Donner zone is not the only zone of local deformation in the range. One other such zone certainly exists west of Dutch Flat, and local deformation probably also occurred along the eastern edge of the Tahoe-Donner trough. When the entire width of the range has been studied in detail the local deformations may be found responsible for all the structural changes in the Tertiary deposits.

Slicing of the range by internal faulting must have greatly reduced its ability to

act as a rigid block throughout its width of 85 miles. If tilting as a single block did occur, it was probably an early event in the Tertiary orogenic history.

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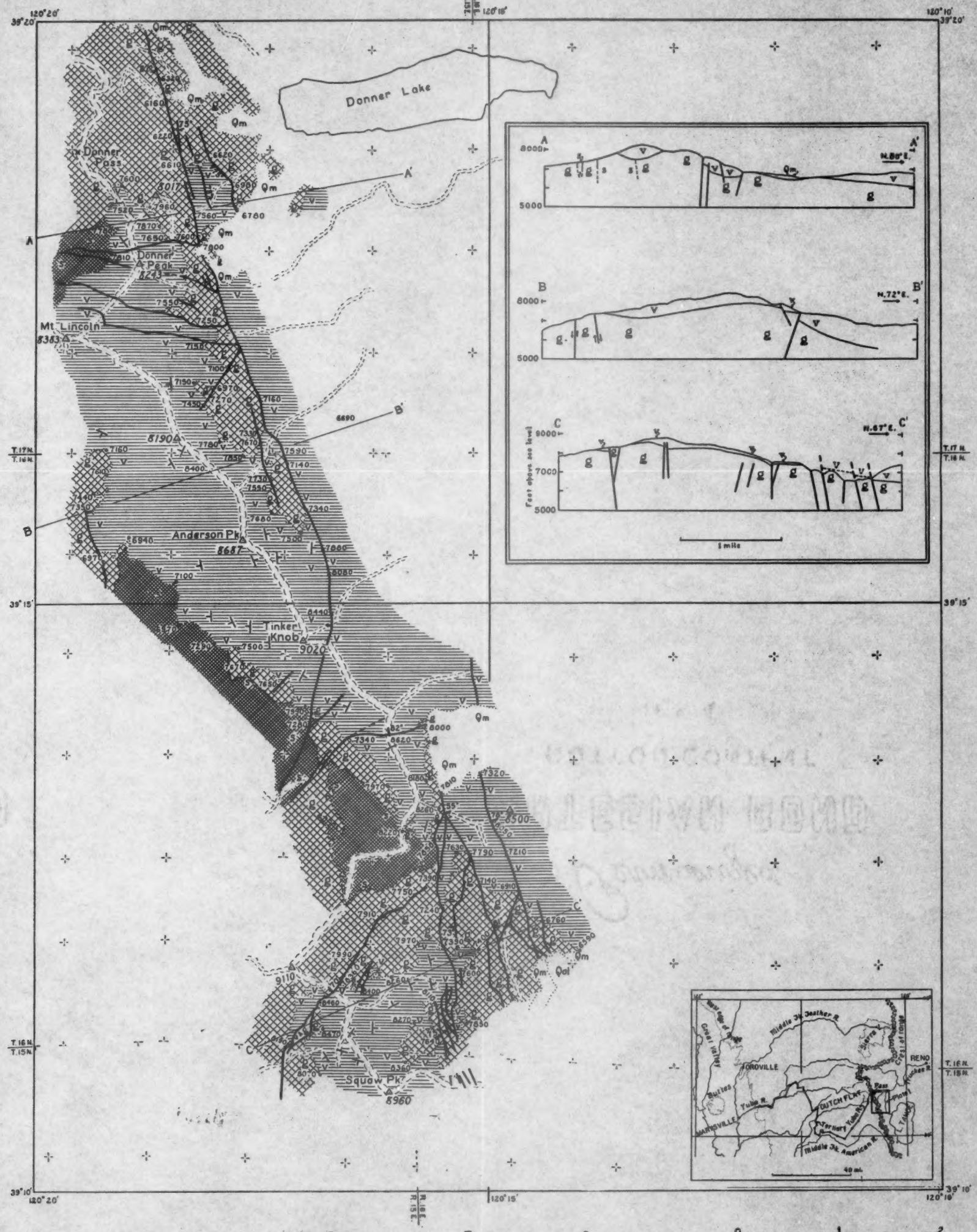
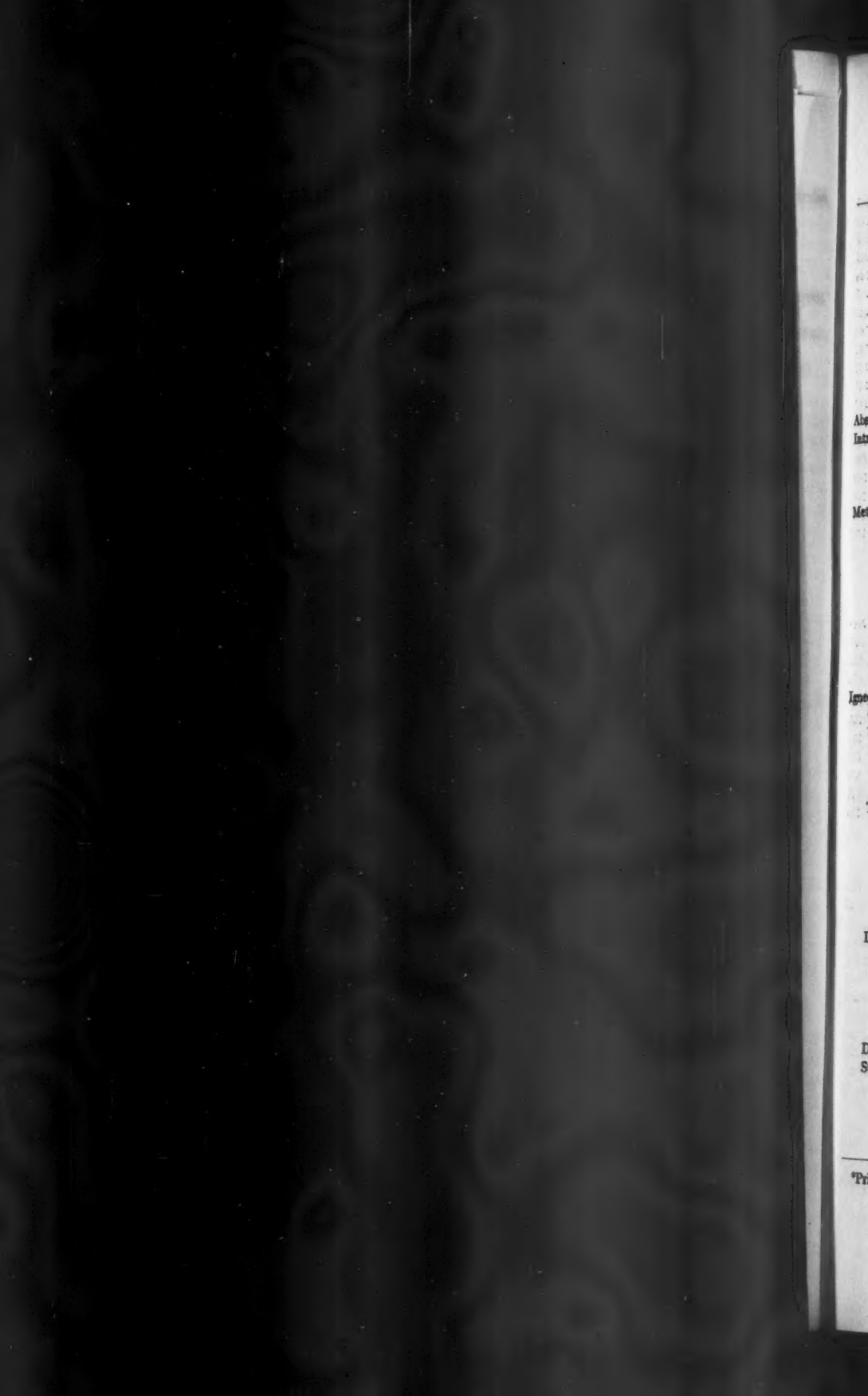



GEOLOGIC MAP AND STRUCTURE SECTIONS, DONNER PASS REGION

With index map of Sierra Nevada in latitude of Donner Pass



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# GEOLOGY OF TOBAGO, BRITISH WEST INDIES*

BY JOHN C. MAXWELL

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*Princeton investigations of Caribbean geology, No. 1.

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## ABSTRACT

Tobago, an island of the British West Indies, lies 20 miles north of Trinidad. The oldest rocks exposed are isoclinally folded schists and phyllites, composed largely of metavolcanic material, outcropping in the mountainous northern part of the island. South of the schist is a belt of strongly sheared but relatively unaltered igneous rocks, including ultramafic and dioritic intrusives, and andesitic and basaltic volcanics. Flat-lying, fossiliferous upper Miocene-Pliocene sands and clays rest unconformably on the volcanics near the southwest end of the island, and coral limestone overlaps both igneous rocks and upper Tertiary sediments.

A review of West Indian geologic history suggests that the Tobago schists are mainly Cretaceous, that the schist-forming movements came in late Cretaceous time, and that a second period of diastrophism, resulting in the shearing of the younger igneous rocks, occurred in the middle or late Eocene. Igneous activity apparently ceased prior to the second diastrophism.

Except for flooding of its margins, Tobago has probably been a land area since late Cretaceous time. During the late Cenozoic, marine sediments accumulated along the present coast line and in drowned stream valleys, subsequently re-excavated. An embayed shore line and prominent sea cliffs indicate recent relative submergence of the island.

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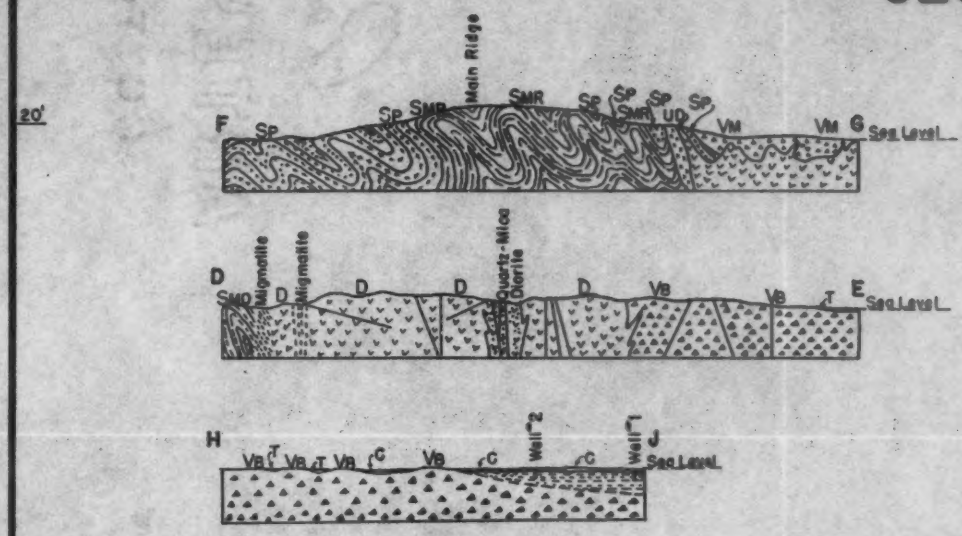
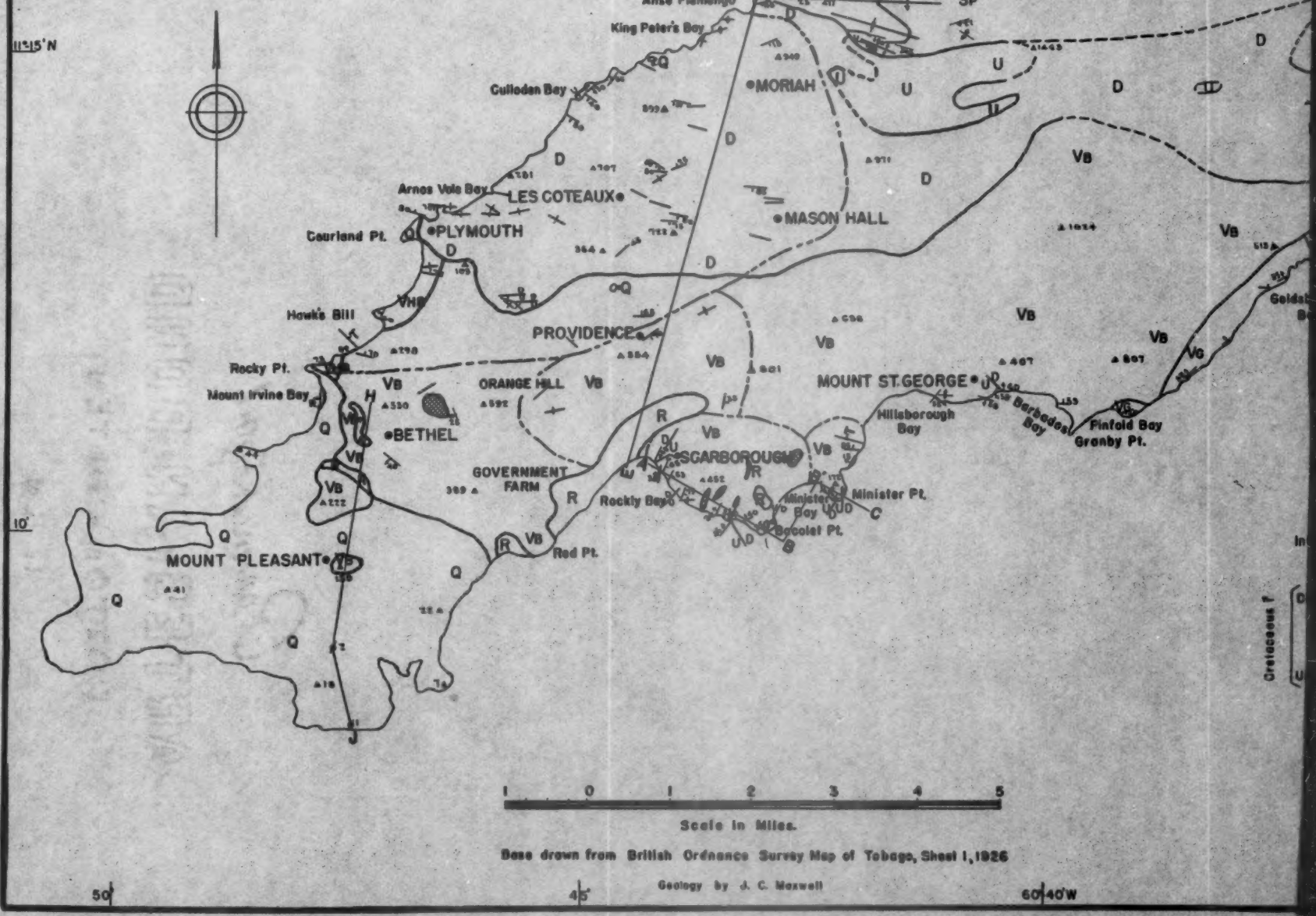
# GEOLOGIC MAP OF TOBAGO B.W.I.




GEOLOGIC MAP OF TOBAGO

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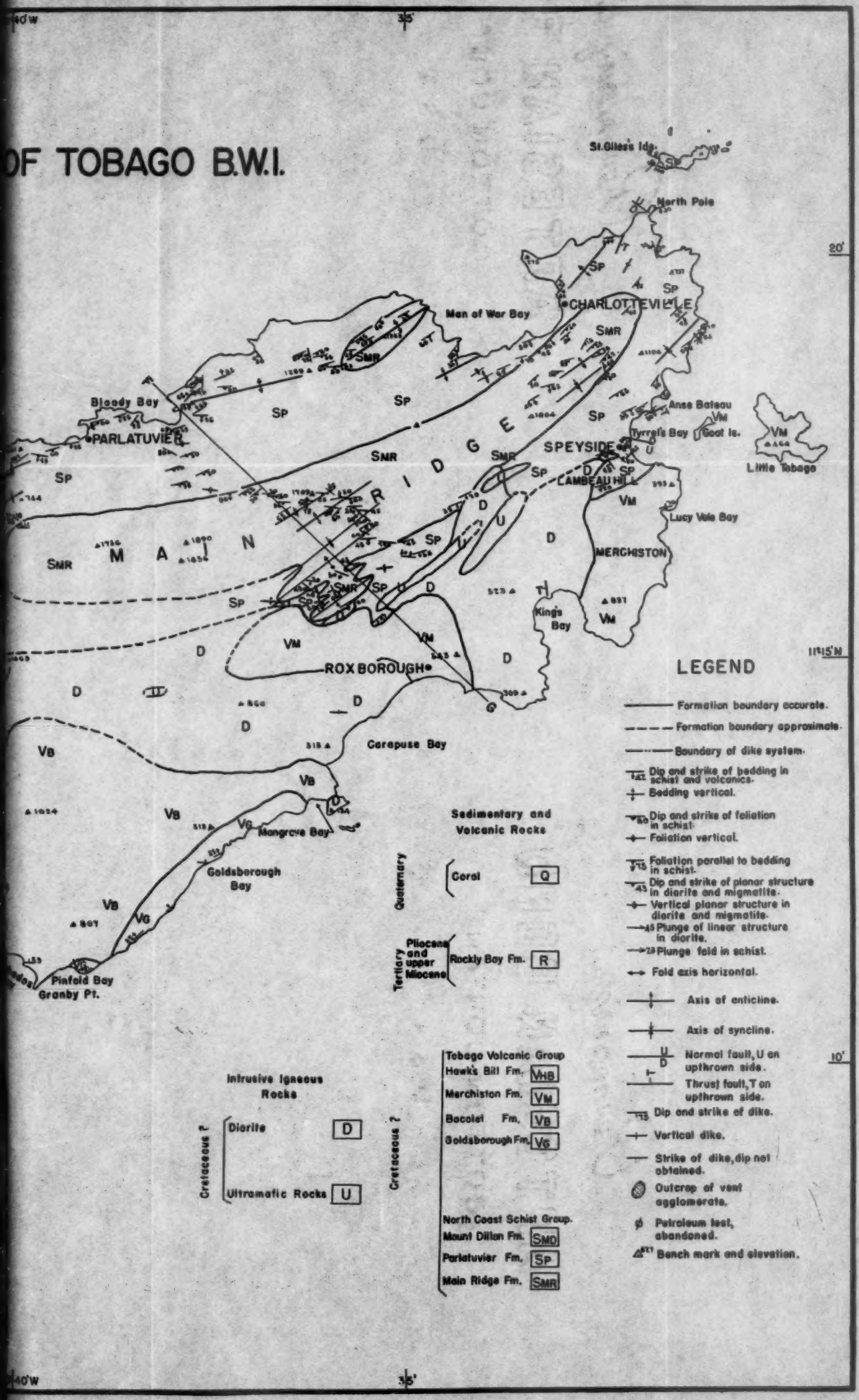


## LEGEND

- Formation boundary accurate.
- Formation boundary approximate.
- Boundary of dike system.
- Dip and strike of bedding in schist and volcanics.
- Bedding vertical.
- Dip and strike of foliation in schist.
- Foliation vertical.
- Foliation parallel to bedding in schist.
- Dip and strike of planar structure in diorite and migmatite.
- Vertical planar structure in diorite and migmatite.
- Plunge of linear structure in diorite.
- Plunge fold in schist.
- Fold axis horizontal.
- Axis of anticline.
- Axis of syncline.
- Normal fault, U on upthrown side.
- Thrust fault, T on upthrown side.
- Dip and strike of dike.
- Vertical dike.
- Strike of dike, dip not obtained.
- Outcrop of vent agglomerate.
- Petroleum test, abandoned.
- Bench mark and elevation.

**Sedimentary and Volcanic Rocks**

Quaternary
- Coral — Q

Tertiary — Pliocene and upper Miocene
- Rockly Bay Fm. — R

Cretaceous ?
- Tobago Volcanic Group
  - Hawk's Bill Fm. — VHB
  - Merchiston Fm. — VM
  - Bacolet Fm. — VB
  - Goldsborough Fm. — VG
- North Coast Schist Group.
  - Mount Dillon Fm. — SMD
  - Parlatuvier Fm. — SP
  - Main Ridge Fm. — SMR

**Intrusive Igneous Rocks**

Cretaceous ?
- Diorite — D
- Ultramafic Rocks — U

OF TOBAGO

# INTRODUCTION

## GEOGRAPHY

Tobago, an island of the British West Indies, lies 20 miles northeast of Trinidad, near the juncture of the Antillean arc and the Venezuelan mainland (Fig. 1). It is

FIGURE 1.—*Location of Tobago*

26 miles long, 7½ miles wide at the widest point, and has an area of 116 square miles. Although the island lies in a humid tropical zone, the southeasterly trade winds give it a pleasant, equable climate. Heavy vegetation and deep weathering characterize much of the island, particularly in the mountainous northeast district where the average annual rainfall may reach 150 inches (Hardy *et al.*, 1931). Other than its fertile soil, Tobago has no known natural resources.

In the northern third of Tobago, deeply eroded schist forms a mountainous area known as the Main Ridge. South of the schist a belt of maturely dissected intrusive and extrusive igneous rocks occupies more than half of the island's area, and at the

southwest end is a low, coral-covered plain. Table 1 summarizes the stratigraphic sequence.

## PREVIOUS WORK

After a geologic reconnaissance of Tobago, Cunningham-Craig (1907) grouped all rocks of the island into two major divisions, metamorphic rocks and Tertiary and post-Tertiary strata. The metamorphic rocks include (1) those of sedimentary and (2) those of igneous origin. Group (1) was described as "schistose grits, very feldspathic, with talcose and talc-mica schists, closely resembling some of the strata in the northern range of Trinidad". He concluded that the original sediments were feldspathic grits, sandstones, and sandy clays. Herein, the rocks of group (1) are named the North Coast schist group and are considered to be largely metavolcanic. According to Cunningham-Craig, group (2) consists of "a mass of basic igneous rock which has shared in the metamorphism of the schists and grits". He mapped it as "epidiorite" and stated that it was intrusive. In the present study, both extrusive and intrusive igneous rocks are recognized, and evidence is presented to show that the igneous material is younger than the schist-forming movements.

Cunningham-Craig dated the Tertiary clays in the vicinity of Scarborough as Pliocene or Recent and thought the coral covering the southwest end of the island was ". . .of quite recent formation".

A sketch map accompanying Cunningham-Craig's paper shows the approximate distribution of the schists, igneous rocks (epidiorite), Tertiary beds, and coral. A small area north of Scarborough mapped as "talcose-quartzose schists" is probably sheared volcanic rocks, and the island of Little Tobago is now known to be composed of volcanic tuffs and tuff-breccias rather than "schists". With these exceptions, the general distribution of lithologic units shown on his sketch map is supported by the present more detailed study.

An excellent discussion of the geography of Tobago accompanies the report of a soil survey by Hardy and others (1931). Trechman (1934) described the Tertiary and Quaternary rocks of the island.

## FIELD WORK AND ACKNOWLEDGMENTS

Field work for this paper was completed during the 3 summer months of 1941. The Department of Geology, Princeton University, provided funds for field and laboratory expenses.

Professor H. H. Hess suggested this investigation, and his constant interest and friendly counsel have aided greatly in its completion. Professors A. F. Buddington, F. B. Van Houten, W. T. Thom, Jr., G. L. Jepsen, and Erling Dorf have read critically portions of the manuscript. Mr. Alan McIlhenny ably assisted in the field.

The geologists of Trinidad Leaseholds, Ltd., Drs. H. G. Kugler, J. C. Griffith, H. H. Suter, and H. H. Renz, extended their hospitality to the writer during two visits to Pointe a Pierre and rendered assistance in many ways. Kugler and Griffith have also given helpful criticism on several parts of the paper.

Dr. Alfred Senn, of the British Union Oil Co., Barbados, spent a day in the field

with the writer and provided suites of rock specimens from St. Barthelemy and Martinique.

It is a pleasure to acknowledge the friendly co-operation of the officials and residents of Tobago during our visit in the trying, early days of the war.

TABLE 1.—*Stratigraphic section*

| Period | Epoch | Formation | | Description |
|---|---|---|---|---|
| QUATERNARY | | Coral | | Coral limestone |
| TERTIARY | Pliocene and upper Miocene | Rockly Bay | | Fossiliferous clays, sands, marls, and conglomerates |
| | | | | —UNCONFORMITY— |
| CRETACEOUS (?) | | Tobago volcanic group | Hawk's Bill | Dacitic lava flows |
| | | | Merchiston-Bacolet | Intrusion of diorite batholith |
| | | | | Andesitic and basaltic tuffs, breccias and agglomerates with minor basalt flows |
| | | | Goldsborough | Dactic tuffs, tuff-breccias, and flows |
| | | | | —UNCONFORMITY— |
| | | North Coast schist group | Mount Dillon | Intrusion of ultramafic rocks |
| | | | | Mica schist, quartzite, and chert with interbedded metavolcanics |
| | | | Parlatuvier | Metavolcanics—greenschists, greenstones, and amphibolites |
| | | | Main Ridge | Phyllitic sericite schists with minor amounts of fine quartzite and metavolcanics |

## METAMORPHIC ROCKS

### NORTH COAST SCHIST GROUP

*General description.*—The North Coast schist occupies the mountainous northern third of Tobago. To the south, the schist is bounded by dioritic and ultramafic intrusives, the line of contact extending from Anse Flamengo on the northwest coast to Tyrrel's Bay on the southeast coast (Pl. 1). Near Anse Flamengo a narrow zone of migmatite marks the boundary, but the eastern portion of the contact zone is a broader belt of contact-metamorphosed schist, intricately intruded by apophyses of diorite.

The North Coast schist group may be divided lithologically into three formations, herein named the Main Ridge, the Parlatuvier, and the Mount Dillon. The oldest of these, the Main Ridge formation, consists predominantly of phyllitic sericite schists which grade upward into a series of metavolvanics, the Parlatuvier formation. Mica schist, thin beds of quartzite and chert, and interbedded metavolcanics characterize the Mount Dillon formation.

*Main Ridge formation.*—Silvery-gray, fissile to platy sericite schist constitutes most of the Main Ridge formation. Minor amounts of fine gray quartzite and metavolcanic greenstone occur interbedded with the schist, the metavolcanics becoming increasingly abundant toward the top of the formation. Much of the schist is stained by hematite produced by weathering of pyrite and chlorite. Where derived from pyrite crystals, the staining is commonly splotchy, but it may also be localized in layers, indicating differences in composition, and hence bedding, in the original shaly sediments. Such evidence of bedding, together with the thin interbedded quartzites, made possible the determination of reliable dips and strikes, in spite of the advanced state of decomposition which weathering under humid tropical conditions imposes on these and, indeed, on all the rocks of Tobago.

As the name indicates, this formation occurs chiefly along the crest of the Main Ridge. The most complete section is exposed along the Roxborough-Parlatuvier track, approximately from the 3½- to the 9⅛-mile marker, and this is designated as the type area of the formation even though inadequate exposures and structural complexity make it impractical to describe the section in detail. By increase in proportion of metavolcanics, the Main Ridge schist grades into the overlying Parlatuvier formation, the boundary being designated as that point in the gradational sequence where metavolcanics become predominant.

A less complete but more accessible section of the Main Ridge formation may be seen on the crest of Back Hill along the Speyside-Charlotteville road, between the 25½- and the 26¾-mile mark. Here also, silvery sericite schist is common, much of it stained by hematite. Metavolcanics make up a considerable proportion of this section, indicating that these are the top-most beds of the formation.

The formation is exposed at only one place near the coast—west of Man-of-War Bay where for a mile or so along the North Side road typical Main Ridge schist may be seen in the low road cuts. Rocks in this vicinity were mapped as talc-mica schists by Cunningham-Craig (1907), as were similar schists elsewhere on Tobago. However, microscopic examination reveals that specimens resembling the "talc schist" of the Hermitage area consist largely of fine-grained white mica. In fact, talc has not been identified with certainty in the Tobago schists, though megascopically many of them appear to be talcose.

*Parlatuvier formation.*—The Parlatuvier formation outcrops almost continuously along the coast from Tyrrel's Bay northward around the end of the island as far west on the north coast as Castara village. Numerous exposures may also be seen along the North Side road from the 10½-mile mark southwest of Castara to Charlotteville, and on the Windward road[1] as far south as the crest of Lambeau Hill. Since the

[1] The Windward road parallels the southeast, or windward, coast of Tobago, connecting Scarborough with Speyside and Charlotteville.

most complete and most readily accessible section of the formation lies along the North Side road and along the north end of the Parlatuvier-Roxborough track, the name of the tiny village of Parlatuvier was given to the formation.

Metavolcanic greenstones and greenschists, with small amounts of interbedded mica schist and quartzite, make up the Parlatuvier formation. In contrast to the other rocks of the North Coast group, the metavolcanics exhibit only a very poorly developed schistosity.

Although the individual rock units of this formation are extremely variable in appearance, they may be grouped lithologically as: (1) faintly schistose, buff, green-speckled rocks with more or less prominent albite porphyroblasts, (2) massive greenstones, (3) hard black massive amphibolites, and (4) fine-grained, bright-green metatuffs.

Rocks of group (1) make up perhaps three-quarters of the Parlatuvier formation. They were called feldspathic grits by Cunningham-Craig, though actually they are porphyroblastic albite-epidote-sericite schists.

Albite occurs both as a constituent of the fine-grained matrix of the schists and as coarse, irregular porphyroblasts, the cleavage faces of which are readily visible to the unaided eye. The porphyroblasts typically have spongy borders and are crowded with inclusions, mostly of clinozoisite, but less commonly of actinolite, chlorite, apatite, or zircon. Many of the albite crystals show evidence of strong movements after their formation. The larger ones are bent or fractured and some have "tails" of finely crushed feldspar trailing off in pin-wheel fashion from opposite corners, indicating rotation of the solid crystal. A few have an hour-glass shape with finely granulated quartz trailing off at either end. The porphyroblasts are either untwinned or relatively coarsely twinned; their composition is constant at $Ab_{92-94}$.

Minerals of the epidote group are abundant in these rocks. Clinozoisite is most common, occurring as faintly pleochroic, colorless to pale-yellow, rounded or prismatic grains, showing low to moderate birefringence, and in some cases anomalous blue interference colors. Dark-yellow epidote may accompany the clinozoisite, and apparently all gradations in composition between the two may appear in the same thin section. Zoisite was identified in only one section.

Of the amphiboles, pale-green or bluish-green actinolite is the most abundant. The optical characteristics of the actinolite (Table 4) are remarkably uniform throughout the entire schist group. Broken fragments of brown hornblende were observed in one specimen. The hornblende is partly altered to greenish-blue actinolite and pale-green chlorite and is probably a relict mineral from the parent igneous material.

The analyzed sample (Tb. 247, Table 2) is fairly characteristic of the buff, feldspathic schists, although it is higher in quartz and lower in amphibole than most of the rocks of this type. Their texture and composition indicate these rocks were derived from basic tuffs, possibly contaminated by varying amounts of fine sand and clay.

*Tb 247* was collected 9⅝ miles northwest of Roxborough on the Parlatuvier track. The sample is pale greenish tan, fine-grained, with a sugary texture. Cleavage faces of feldspar porphyroblasts give the rock a sparkling appearance. The feldspar is albite ($Ab_{94}$); it shows coarse albite and Carls-

bad twinning but is unzoned. Albite porphyroblasts have ragged margins and are crowded with inclusions of clinozoisite, quartz, and actinolite. A few porphyroblastic grains of quartz are present, likewise containing abundant inclusions of clinozoisite and actinolite. The larger quartz grains are strained, and their edges granulated. The groundmass is chiefly clinozoisite, probably with a considerable amount of sericite, though this mineral was not positively identified. Scattered patches of

TABLE 2.—*Chemical analysis, norm and mode of albite-clinozoisite schist*

R. B. Ellestad, Analyst

| Chemical Analysis | | Norm | | Mode | |
|---|---|---|---|---|---|
| | | Weight per cent | | | |
| $SiO_2$ | 57.99 | Qtz | 13.74 | Quartz | 3.8 |
| $Al_2O_3$ | 18.90 | Or | 2.78 | Albite | 42.0 |
| $Fe_2O_3$ | 4.49 | Ab | 36.68 | Clinozoisite (+ sericite) | 51.0 |
| FeO | 0.91 | An | 31.69 | | |
| MgO | 1.20 | Di | 7.56 | Serpentine | 2.2 |
| CaO | 8.80 | Ilm | 1.52 | Amphibole | 1.0 |
| $Na_2O$ | 4.29 | Mt | 0.93 | | |
| $K_2O$ | 0.48 | Apt | 0.34 | | |
| $H_2O^+$ | 1.57 | Wo | 0.93 | | |
| $H_2O^-$ | 0.23 | Hem | 4.00 | | |
| $CO_2$ | trace | | | | |
| $TiO_2$ | 0.83 | (Pc = $Ab_{21}$ $An_{79}$) | | (Pc = $Ab_{94}$ $An_6$) | |
| $P_2O_5$ | 0.16 | | | | |
| MnO | 0.10 | | | | |
| Totals | 99.95 | | 100.17 | | 100.0 |

a honey-yellow mineral with a faintly fibrous structure, low birefringence, and $\beta = 1.634$ (bowlingite?) together with minute feldspar and quartz grains make up the remainder of the groundmass.

Group (2), the greenstones of the Parlatuvier formation, includes hard, pale-green, schistose to gneissic metavolcanics in which the green color is due to epidote or chlorite, actinolite being uncommon. Albite porphyroblasts may constitute as much as 60 per cent of the rock, or they may be absent. Where abundant, the porphyroblasts stand out prominently on weathered surfaces, giving the rock much the appearance of a pebbly grit. Quartz and calcite are common in some of the greenstones, absent in others. Their appearance in some specimens suggests derivation from amygdaloids in the parent igneous rocks.

*Tb 245*, a stream pebble from Bloody Bay River, is an example of a greenstone which has retained amygdaloidal texture. The rock, containing scattered elliptical quartz bodies 5 to 20 mm. in diameter, closely resembles an amygdaloidal basalt, except that the groundmass is faintly schistose and yellow green. Microscopically the groundmass consists of a felt of irregular, twinned feldspars and tiny subhedral grains of epidote, with a few larger porphyroblastic crystals of epidote. The feldspar is apparently porphyroblastic also, since it is crowded with inclusions of epidote and actinolite. The quartz bodies are made up of interlocking grains showing wavy extinction. Many of the bodies have rims of felted actinolite and epidote, the latter being yellowish green and much darker than epidote of the groundmass. Presumably the rim of dark minerals around the quartz corresponds to the chlorite-epidote rim frequently observed around amygdules in basalt. Some migration of quartz evidently took place during recrystallization of the rock, for quartz veinlets extend outward from some

of the larger "amygdules". Quartz grains in the veinlets are strained, as in the larger bodies, but the veinlets do not show a concentration of dark minerals along their walls.

Rocks of group (3), hard, black, massive amphibolites, occur at only a few places within the schist. Apparently they were formed from andesitic flows or sills during the regional metamorphism. Near the diorite-schist contact are rocks of similar appearance, but these are definitely a contact-metamorphic facies and are discussed separately.

*Tb* 254 from Little Englishman's Bay is typical of group (3) rocks. Megascopically it resembles a fine-grained melanocratic igneous rock, with just a suggestion of gneissic structure. Under the microscope, however, the rock is seen to be a hornblende-albite gneiss, consisting essentially of an aggregate of anhedral albite prophyroblasts with interspersed whisps and needles of amphibole and lesser amounts of chlorite and epidote. Albite crystals (about $Ab_{98}$), showing coarse albite twinning but no zoning, are crowded with inclusions of epidote, hornblende, and apatite. Optically the amphibole is an actinolitic hornblende, with (−) 2V = about 50°, $Z \wedge c$ = 19°, X = pale yellow, Y = yellowish green, Z = pale blue. The chlorite is probably clinochlore. It has (+) 2V = about 10° and is pleochroic, X = pale olive green, Y = pale yellow green, Z = light green. Epidote occurs as tiny yellow subhedral grains as well as inclusions in the albite porphyroblasts. A few grains of zoisite were also observed. Scattered irregular grains of quartz showing wavy extinction intertongue with the albite. Abundant subhedral to anhedral pyrite metacrysts are developed in nests of femic minerals and appear to replace these minerals. In its field relations this hornblende-albite gneiss body is a 90-foot thick layer conformable to the adjacent schists.

Group (4), the fine-grained, bright-green metatuffs, are most abundant near the Speyside Hotel on the north side of Tyrrel's Bay. These hard, massive, finely granular rocks consist of interlocking grains of anhedral epidote, feldspar, and minor quartz, and many contain large, unoriented porphyroblasts of black amphibole. Such porphyroblasts, up to 2 cm long, occur in a sample (Tb 293) collected from the Speyside Hotel beach. In thin section, the amphibole is pale-greenish yellow, very faintly pleochroic, with (−) 2V about 80° and $Z \wedge c = 18°$. Much of it has been altered to colorless, nearly isotropic chlorite and colorless tremolite. The tremolite develops parallel to the orientation of the hornblende but occurs only in the completely chloritized portions of the porphyroblasts.

The mineral association, fineness and uniformity of grain of the groundmass and the unusual development of euhedral porphyroblasts suggest derivation from a permeable, basic vitric tuff. Inasmuch as the porphyroblasts are unoriented and are not distorted or crushed, they probably developed after the regional metamorphism, aided by emanations from the near-by diorite. Presumably the alteration of the hornblende porphyroblasts to chlorite and tremolite resulted from continued emanations acting at decreasing temperatures.

*Mount Dillon formation.*—The Mount Dillon formation is found only at the extreme western end of the Tobago schist outcrop. Excellent exposures may be seen along the Moriah-Castara road, skirting the western flank of Mount Dillon. Beginning at about the 8⅛-mile marker, typical Mount Dillon schists are exposed almost continuously to the contact with the Parlatuvier formation near the 10½-mile marker. This is designated the type area of the formation.

The Mount Dillon formation consists of white to buff mica schists, thin-banded, fine gray quartzites, cherty beds, and a little graphitic schist. Soft, buff metatuff

beds similar to group (1) of the Parlatuvier formation make up a considerable part of the section. The cherty layers interbedded with white mica schist show no trace of organic structures.

Near the 9½-mile road marker a 15-foot long, downfaulted block of graphitic or carbonaceous schist was observed. The schist is dark gray to black, deeply weathered and soft, and resembles a carbonaceous shale. No other outcrop of such material was noted.

*Metamorphic facies.*—The following mineral assemblages characterize the rocks of the North Coast schist group: (1) Albite-epidote (clinozoisite)-actinolite-chlorite-sericite-(quartz); (2) Albite-clinozoisite-actinolite; (3) Albite-clinozoisite (epidote)-chlorite-calcite-(quartz); (4) Epidote-chlorite-actinolite-(quartz); (5) Epidote-clinozoisite-actinolite.

The first two make up the bulk of the schist group; 3, 4, and 5 are comparatively rare.

The association of chlorite with sericite, and of albite with epidote, indicates that these rocks belong to the greenschist metamorphic facies of Eskola (1920). The work of Turner and others would seem to place these schists in the higher metamorphic grades within the greenschist facies. Turner (1941) recognized a chlorite zone in the schists of Otago, New Zealand, which is apparently about equivalent to Eskola's greenschist. Within this zone he differentiated four subzones representing stages of increasing metamorphism, of which the highest, subzone chlorite 4, is a "...completely reconstituted, thoroughly laminated coarse-grained schist, in which the average grain size for albite and quartz typically lies between 0.25 and 2 mm. (p. 3)." Most of the North Coast schists fit this description.

Likewise Harker (1939, p. 280) stated that progressive metamorphism within the chlorite zone is marked by the gradual replacement of chlorite by hornblende. In the schists of Tobago, actinolitic hornblende is preponderant over chlorite, though both may be present in the same rock.

Hutton (1940) has shown that the epidote mineral changes markedly with increasing grade of metamorphism, the yellowish granular variety characterizing the lower grades while the more highly metamorphosed rocks carry colorless, idioblastic prisms of clinozoisite. Most of the Tobago schists are near the clinozoisite end of this sequence, but apparently equilibrium is not always attained, for yellowish granular epidote is found in some schists to the exclusion of clinozoisite, and in others yellow epidote and colorless clinozoisite occur in the same specimen.

Turner (1935, p. 417) stated that actinolite-albite-epidote-chlorite schists may be formed by dynamothermal metamorphism at high stress and relatively low temperature. Probably such conditions prevailed during the metamorphism of the North Coast schists. Where higher temperatures were induced near the intrusive diorite mass, rocks of the greenschist facies, or chlorite zone, were converted to the next higher grade of metamorphism, the amphibolite facies of Eskola (1920, p. 155), or the biotite zone of Harker (1939, p. 214).

*Schistosity.*—Rocks of all three formations within the Schist group characteristically show bedding-plane schistosity. At the type section of the Mount Dillon formation, for example, quartzite beds in micaceous schist emphasize this type of schistosity, and within the Main Ridge formation similar relations may be noted along the Rox-

borough-Parlatuvier road. Even in the relatively massive Parlatuvier metavolcanics, field data reveal a definite parallelism of schistosity and bedding. At a few outcrops only could schistosity be seen to cut the bedding, and here the dips are generally erratic, suggesting that both features are related to the axes of minor folds.

Variations in degree of development of schistosity in the metavolcanics and the metasediments are attributed to differences in composition of the parent rocks rather than to differences in tectonic history. In the metasediments, the conspicuous cleavage is due to abundant mica. In the metavolcanics, on the other hand, mica is not common, and the imperfect schistosity is due largely to a planar orientation of stubby actinolite and epidote crystals or, in a few rocks, to the development of chlorite.

*Structure.*—The outcrop pattern (Pl. 1), attitude of bedding, and minor folds indicate that the North Coast schists form an anticlinorium composed of isoclinal folds which are overturned toward the north or northwest (Pl. 1, cross section F–G). The folds plunge to the west in the vicinity of Mount Dillon and, from the areal geology, would seem to plunge northeastward at the east end of the island, a supposition supported by the presence on St. Giles island of a small anticline plunging gently northeast.

Within the Mount Dillon formation, drag folding in mica schist and fracture cleavage crossing quartzite beds made possible a fairly accurate determination of minor fold axes. Correlation of quartzitic horizons within the formation gave supporting evidence, though the correlations alone could not be considered reliable. Areal geology and dip and strike data reveal the sharp, plunging folds 1½ miles northwest of Roxborough. West of Man-of-War Bay, the exposure of Main Ridge sericite schist suggests the presence of an anticline. Other fold axes were located on the basis of erratic dips and strikes and on evidence as to the attitude of bedding (upright or overturned) furnished by exposures exhibiting drag folding, cross-cutting schistosity, or fracture cleavage. Because of the difficulties of mapping, the delineation of minor folds is incomplete and probably must remain so.

No evidence of major faulting involving displacements of the order of hundreds or thousands of feet has been found in Tobago, but minor faulting is common in both the igneous and metamorphic rocks. That in the igneous rocks is discussed in a later chapter. Within the schist group, both normal and reverse faults were observed, principally in sea cliffs where large, fresh surfaces are exposed. In the deeply weathered schist exposures found inland faulting would be extremely difficult to detect. Reverse faulting, with the fault plane striking northeast but oblique to regional strike, was observed in the small bay southwest of North Pole and on the headland between Tyrrel's Bay and Anse Bateau. In both cases overthrusting from the southeast is indicated.

*Thickness.*—Because of structural complexity, the thickness of the North Coast schist could not be determined accurately. The available data suggest a minimum of 7800 feet.

Cross section F–G (Pl. 1) reveals that a 1200-foot thickness of Main Ridge schist is needed to account for the outcrop width of that formation, assuming that folding is no more complex than shown. Since the base of the formation is not exposed, this is a minimum figure.

The entire thickness of the Parlatuvier formation is present near the west end of

the North Coast schist outcrop. Unfortunately, its contact with the Main Ridge schist lies in densely wooded country and could not be accurately located. A thickness of 3100 feet has been estimated. In view of the large outcrop area of the formation, this figure does not seem excessive.

To obtain the thickness of the Mount Dillon formation, it was necessary to measure the beds along the crest of the westward-plunging folds. Drag folds indicate that the angle of plunge increases from 23° near the eastern margin of the outcrop to 50° at the coast. These structural conditions require a thickness of 3500 feet of sediments; the figure seems excessive, but no better estimate can be given.

## IGNEOUS ROCKS

### GENERAL STATEMENT

South and southeast of the North Coast schist, younger igneous rocks outcrop over more than half of the island's area. The oldest of these is an ultramafic series, followed by extrusive and intrusive rocks of the Tobago volcanic group and by a complex dioritic intrusion of batholithic proportions. Andesitic and dacitic dikes, and possibly dacitic lava flows, appear to be the youngest igneous rocks.

### ULTRAMAFIC INTRUSIVES

*General description.*—The ultramafic rocks occur for the most part as small, lenticular bodies near the diorite-schist contact, bounded on one side by diorite and on the other by schist. In a few places, however, the ultramafics are wholly within schist or diorite (Pl. 1). The largest, and westernmost, ultramafic mass measures about 2½ miles in length and a mile in width. Those to the east are somewhat smaller. All known ultramafics are shown on Plate 1, with the outcrop areas of the smaller bodies necessarily somewhat exaggerated.

Coarse-grained augite peridotite predominates among the ultramafic rocks exposed on Tobago, serpentinite is less common, and pyroxenite relatively rare. The massive, black peridotite and pyroxenite weather to a characteristic olive-green saprolite in which the texture of the parent rock is preserved. Unweathered material is found most commonly as residual boulders within the saprolite.

The serpentinite is a fine-grained, dense, black rock, complexly sheared and slickensided. It occurs both as isolated masses and as irregular bodies associated with peridotite. Examples of isolated masses may be seen at the head of Kendall road, 2¼ miles northwest of Roxborough, and near the east edge of Speyside village. In the three lenticular ultramafic bodies outcropping northwest of King's Bay, the serpentinite apparently passes laterally into peridotite, while in the large mass near the west coast serpentinite occurs as thin sheets or lenses in coarse peridotite. Where they occur together, no particular pattern was recognized in the distribution of serpentinite relative to peridotite.

Peridotite and pyroxenite occurring near the margins of ultramafic bodies show strong shearing, probably related to post-intrusion, regional deformation. The pyroxene crystals are crushed and deformed, but not chemically altered. On the other hand, the small amount of olivine present in all the peridotite has been more or less

serpentinized, and serpentinite is itself associated with both sheared and unsheared peridotite. Apparently the process of serpentinization bears no relationship either to marginal shearing or to solutions emanating from the adjacent diorite batholith. Rather, the alteration of the olivine to serpentine probably occurred during the last stages of intrusion of the ultramafic rocks, as postulated by Benson (1918) and Hess (1933) for the majority of serpentinites.

*Age and mode of intrusion.*—The ultramafic rocks probably are younger than the North Coast schists, but older than the diorite and volcanics. The ultramafics are intruded by dikes of diorite and, where adjacent to the diorite batholith, show the effects of contact metamorphism. Dikes of diorite and diorite pegmatite were observed in the ultramafics at three localities. On the Mena road, 1½ miles north of the main road extending northeast from Mason Hall, small dikes of medium-grained meladiorite occur in deeply weathered peridotite. At the second exposure, about 6⅛ miles north of Mount St. George on the Castara road, dikes of light-colored feldspar-hornblende pegmatite cut coarse peridotite. The third locality, 1¾ miles north of Roxborough on the Parlatuvier track, shows sheared pyroxenite intricately intruded by dikes of coarse meladiorite and light-colored, feldspar-hornblende pegmatite. A highly pegmatitic facies of meladiorite outcrops for a quarter of a mile south of the pyroxenite, and to the south of this are found lapillituffs, also intricately intruded by apophyses from the diorite body. The pyroxenite, diorite, and volcanic tuffs all show intense shearing and small-scale faulting.

In a thin section of sheared pyroxenite (Tb 259) from the third locality, the pyroxene, an augite (Table 4), shows partial alteration to semiopaque, dusty material, the alteration being localized somewhat along a prominent basal parting. Cloudy, twinned albite occurs sparingly in the rock, interstitial to augite. Shearing movements subsequent to the alteration caused bending and fracturing of the augite crystals and the development of mylonite zones containing fragments of albite and augite.

The introduction of albite and slight alteration of the pyroxene was probably accomplished by solutions from the diorite. Subsequent mechanical deformation, in which the diorite and near-by volcanic rocks participated, produced the mylonite zone and the distortion of the pyroxene crystals. The same deforming forces probably caused the shearing noted in the marginal rocks of all the ultramafic bodies, as well as the widespread and intense shearing present throughout the diorite and volcanics.

Evidence for an intrusive relationship between the schist and ultramafics is indirect, for no definitely intrusive contact between the two was found, and the schists adjacent to ultramafic bodies show no detectable contact metamorphic effects. However, the geologic map shows that the ultramafics follow no one stratigraphic horizon, as would be expected if they represent in-folded basement rocks. Near the west coast, ultramafics are in contact with the highest beds of the Parlatuvier formation, while the small outcrop of serpentinite northwest of Roxborough lies about at the contact between the Parlatuvier and Main Ridge formations. Other ultramafic bodies occur at various horizons within the Parlatuvier formation or within the diorite.

Absence of contact metamorphic effects in schist adjacent to the ultramafic rocks is not conclusive evidence against an intrusive relationship between the two. Many writers have pointed out a similar lack of metamorphism around intrusive ultramfic bodies. Hess (1938, p. 325) described peridotite lenses 200 feet thick and 1000 feet long showing no obvious contact effects on argillaceous sandstones. Du Rietz (1935, p. 138) noted a similar lack of contact metamorphism around small peridotite masses in Sweden, as does Vermunt (1937, p. 18) in Cuba. In the Mt. D'Or area of New Caledonia, where an ultramafic body of batholithic proportions has intruded soft sandstones and shales, the writer observed only a slight serpentinization of sandstone and baking of shale within 30 yards of the contact. Sediments a few yards farther away are unaltered.

Wherever the structure of the schist adjacent to ultramafics could be determined, the ultramafics lie in or near the crests of anticlines—for example, the northwest edge of the westernmost ultramafic body, the small serpentinite body northwest of Roxborough, and the larger of the two bodies near Tyrrel's Bay (Pl. 1). In the last-named example, the peridotite occupies an irregular domal structure. Thayer (1942, p. 7) noted that serpentine in Cuba is exposed in the central portions of domal or anticlinal uplifts, and Du Rietz (1935) described small ultramafic bodies related to anticlinal folds in northern Sweden. Du Rietz considered the ultramafics to be hypabyssic intrusions, usually in the form of small laccoliths (1935, p. 139). Serpentine lenses in the Graipsvare area he regarded as tectonic intrusions in anticlinal ridges, possibly derived from flat laccoliths or sheets which were broken and up-ended during the regional folding. Most of the rocks Du Rietz described are dunites or olivine-rich peridotites showing intense serpentinization, shearing, and slickensiding. The less abundant peridotites, he assumed, would slide easily because they are also largely composed of serpentine.

In the Tobago ultramafics, on the other hand, serpentine is not abundant, the major constituent being fresh peridotite containing only a few per cent of serpentinized olivine grains. On the margins of the larger bodies the peridotite is, indeed, strongly sheared, but the centers are invariably massive and show no evidence of deformation. Had the peridotite bodies been injected tectonically by solid flow, they should show more intense shearing throughout. Most likely, the ultramafics were intruded into the schists at moderate depths after the fold pattern had been well developed. The sheared margins of the bodies may indicate a continuation of the folding after injection of the igneous material, or, as pointed out above, the shearing may have been related to the later movements which affected the younger diorite and volcanics.

No contact was observed between ultramafics and the pyroclastic rocks of the Tobago volcanic group. The volcanics are thought to be the younger, however, for they apparently are not intruded by the ultramafics. Also, the pyroclastic rocks lie unconformably on schist and so post-date the probable period of ultramafic intrusion.

*Petrography.*—Augite and olivine, plus serpentine derived from olivine, are the common minerals of the ultramafic rocks. A very small amount of primary magnetite is present in all sections, and secondary magnetite is abundant in serpentine. Rhombic pyroxene and primary hornblende and feldspar seem to be absent, though

locally uralite developed and albite was introduced during contact metamorphism by diorite. Chromite was not identified in any of the sections;[2] however, McIlhenny (1942) found chromite octahedra in beach sand from Hillsborough Bay.

The peridotites and pyroxenites are coarse-grained, dark-green to black rocks with hypautomorphic-granular textures; they are composed of medium- to dark-green augite and dull black serpentine-magnetite derived from olivine. No fresh olivine can be detected in hand specimens, though it is present in many thin sections. The peridotites contain from 60 to 90 per cent augite, the remainder being olivine, or serpentine and magnetite derived from olivine. Of the two pyroxenites from which thin sections were prepared, one was free of olivine, the other carried a little dark serpentine, probably indicating a few per cent of olivine in the original rock. Optical constants of minerals from peridotite and pyroxenite are listed in Table 4.

The serpentinites are dull-black, fine-grained rocks, intricately sheared and slickensided, derived from dunites or olivine-rich peridotites. The olivine of the dunite (Tb 330) is represented by pale grayish-green antigorite with characteristic mesh structure. Minute veinlets of colorless, fibrous serpentine are present, and the rock shows limonite staining along irregular, branching cracks which apparently antedate the formation of the chrysotile veinlets. Magnetite occurs as discrete grains and as disseminated dust and veinlets in the serpentine.

Olivine of the olivine-rich peridotite (Tb 203) has been altered to greenish-yellow to colorless antigorite, though the xenomorphic-granular texture of the rock persists. Olivine pseudomorphs make up about 80 per cent of the rock, the remainder being large grains of dark brown, nearly isotropic bastite poikilitically enclosing olivine pseudomorphs. The bastite shows parallel extinction, suggesting it was derived from orthorhombic pyroxene. The unaltered rock would, therefore, be classed as a saxonite. Secondary magnetite is concentrated around grain boundaries of the olivine pseudomorphs and along relict pyroxene cleavages in the bastite.

Wherever diorite is adjacent to ultramafic rocks, the latter show evidence of contact metamorphism, although the changes may be recognizable only in thin section. The introduction of albite into pyroxenite and partial alteration of augite to a dusty, opaque substance has been described previously. More commonly the pyroxene has been partially altered to pale brownish-green or bluish-green hornblende, and epidote introduced as cross-cutting veinlets. Only near diorite does hornblende occur in peridotite and pyroxenite. For description of minerals see Tb 194, Table 4.

A small "serpentinite" body exposed at the east edge of Speyside village has apparently undergone both contact metamorphism and subsequent mechanical deformation. Megascopically the rock (Tb 309) has the dull-black to greenish-black appearance of a normal serpentinite, but is more intensely sheared and is criss-crossed by veinlets of a pale-green mineral. The thin section shows the rock to consist of a mass of pale-bluish-green, pleochroic actinolite (Table 4) crossed by veinlets of finely granular epidote. A network of minute fractures and shear planes displaces the epidote veins, and the actinolite has a cataclastic texture. The original rock was probably peridotite, into which epidote was introduced during contact metamor-

[2] The "ore mineral" is black, strongly magnetic, and completely opaque on thinnest edges of grains. Bead test with phosphorus salt of black opaques from four samples (TB 194, 205, 311, 330) showed only iron, no chromium nor titanium.

phism, the olivine and augite being replaced by actinolite at the same time. Subsequent deformation produced the shearing and cataclastic textures. During the metamorphism the nearby schist was altered to an amphibolite, and also intensely sheared.

Augite peridotite, the predominant ultramafic rock of Tobago, apparently is not a common rock type. Wehrlites described by Turner (1930) from New Zealand may be similar, though these rocks average 70–80% olivine, and carry accessory chromite and a little enstatite. Nockolds (1940, p. 452) described augite peridotite from Scotland which is similar to the Tobago rocks. Judging from optical data, however, the augites of the two series have somewhat different compositions. Nockolds gives the following data for an analyzed augite: $\alpha = 1.680$, $\beta = 1.686$, $\gamma = 1.705$ ($\pm.003$), $2V = 61°$, $Z \wedge c = 46°$. The indices are close to those of the augite from Tobago (Table 4), but 2V and $Z \wedge c$ are both several degrees larger. In Scotland, as in Tobago, brown hornblende developed from the augite of the peridotites and pyroxenites, probably as a result of contact action by later diorite or gabbro (Nockolds, p. 455). This hornblende differs somewhat in optical properties from the corresponding Tobago hornblende, as should be expected since the parent augites differ in composition.

## TOBAGO VOLCANIC GROUP

*General statement.*—Rocks of the Tobago Volcanic Group outcrop over much of the southern and southeastern parts of the island (Pl. 1). On the basis of distribution and lithologic characteristics, the volcanics may be divided into four formations. The oldest of these, called the Goldsborough formation, consists of pale greenish-gray dacitic tuffs, breccias, and flows. It is overlain by andesitic and basaltic pyroclastic rocks of the Bacolet formation; similar rocks outcropping to the north, on Merchiston peninsula and Little Tobago island, have been designated the Merchiston formation. Dark dacite flows assigned to the Hawk's Bill formation lie on Bacolet tuffs near the southwest end of Tobago.

*Terminology.*—Unless otherwise noted, the terminology and classification of pyroclastic rocks used in this paper is that recommended by Wentworth and Williams (1932). The terms "welded tuff" and "welded tuff-breccia" will be used for those rocks which have the textures of tuffs or tuff-breccias but which have become indurated at high temperatures by the "welding together" of glassy fragments.

*Goldsborough formation.*—The Goldsborough formation outcrops in a narrow strip along Goldsborough Bay, the type area for the formation. Their pale greenish-gray color and pinkish feldspar phenocrysts distinguish these rocks from the dark-green to black pyroclastics of the adjacent Bacolet formation.

Near the southwest end of the outcrop, tuff-breccia and fine agglomerate predominate. These are hard, well indurated rocks, cut by closely spaced joints which give a jagged, splintery appearance to the exposures. Dacite blocks in the agglomerate and tuff-breccia average 2 to 4 inches in diameter and are rounded to subangular, of aphanitic or glassy texture, and may show flow banding and chalcedonic amygdaloids. The interbedded flows, from 1 to 12 ft. thick, are light- to dark-gray aphanites.

In Goldsborough Bay and to the northeast, the coarser pyroclastics give way to

fine tuff-breccias and tuffs. These also are hard, splintery, and pale greenish gray in color. North of Goldsborough Bay, however, the tuffs are intricately intruded by dikes and sills of diorite which have metamorphosed the adjacent volcanics to dark-gray hornblendic rocks (Pl. 2, fig. 1).

The age of the Goldsborough volcanics is not known. They are almost certainly older than the Bacolet formation, for they dip 35° to 55° northwestward beneath Bacolet tuffs. Although the steep dips imply post-depositional disturbance of the volcanics, the deformation has been primarily of the nature of steepening of original dips rather than of strong folding. In the absence of evidence for overturned folds, it may be assumed that the Goldsborough rocks are older than those of the Bacolet formation. There may be some overlap in ages, however, for a 15-foot thick layer of black, pyroxene-porphyry tuff-breccia is interbedded with dacite agglomerates about half a mile southwest of Goldsborough Bay. Dark pyroxene-bearing tuffs also occur in the sea cliff just north of Pinfold Bay, dividing the outcrop of the dacitic volcanics, but this occurrence is interpreted as a valley-filling on the surface of the Goldsborough volcanics.

The source of the dacite lavas and pyroclastics is unknown. No vent agglomerates or coarse pyroclastics are found to indicate proximity to the old volcano. Nevertheless, the very constant northwest dip of all beds and the occurrence of agglomerates south of Goldsborough Bay point to an offshore source, located somewhere to the southeast of the agglomerate outcrops.

*Bacolet formation.*—Except for small areas on the northwest coast at Hawk's Bill and on the southeast coast in the vicinity of Goldsborough Bay, the Bacolet formation includes all the volcanic rocks south of the diorite batholith. The formation is composed of interbedded tuffs, tuff-breccias, and agglomerates, with a very few intercalated flows. Crosscutting dikes are abundant, and pipelike bodies of vent agglomerate and intrusive volcanic breccia are locally present. Excellent exposures of the rocks of this formation occur along the sea coast in the vicinity of Bacolet Point, which is designated the type locality of the formation.

The coarsest tuff-breccias and agglomerates occur within a narrow coastal belt extending a mile or so inland and about 3 miles to either side of Bacolet Point. Most of the occurrences of intrusive agglomerate and breccia are found in this same belt. Finer tuff-breccias and tuffs, nearly everywhere deeply weathered, cover the inland area but are also found interbedded with the coarser pyroclastic rocks near Bacolet Point.

Figure 2 shows, somewhat diagrammatically, the interbedded character of the pyroclastic rocks. Actually, the rocks nearly everywhere appear to be massive and unstratified with gradations in average grain size so gradual as to be almost unnoticeable within any one exposure. Thin beds of fine tuff and the basal contacts of agglomerate with underlying finer material are the only commonly observed indications of stratification.

The agglomerates typically consist of rounded to subangular blocks of amygdaloidal basalt and andesite porphyry in a matrix of brownish-green, easily weathered tuff. Ranging in size from lapilli to masses 10 feet or more across, the blocks average about 2 to 4 inches in diameter (Pl. 2, fig. 2).

In the coarsest agglomerate observed, subangular blocks 10 feet or more in diameter rest in a tuff matrix in the sea cliff below Bacolet House. The large blocks are themselves agglomeratic, being made up of smaller blocks and lapilli of greenish-black pyroxene and feldspar porphyry in a welded tuff matrix. West of Bacolet beach, somewhat less coarse agglomerate contains blocks a foot or so in diameter, not

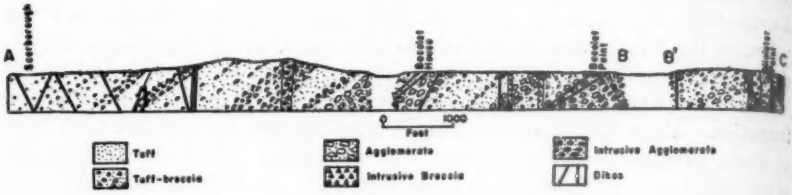


FIGURE 2.—*Section A–B–B¹–C, Bacolet area*

agglomeratic, yet varying considerably in petrographic character. Most are porphyritic, with greenish-black pyroxene porphyry predominating. Dark-gray feldspar porphyry and black amygdaloidal basalt blocks are also common. This heterogeneity of rock type is characteristic of the agglomerates and coarser tuff-breccias of the Bacolet volcanics.

Agglomerates of the Bacolet area grade upward (northwestward) into finer-grained pyroclastic rocks, as is well shown in the sea cliff southeast of Scarborough (Fig. 2). Eastward from Minister Point a similar gradation from coarse to fine material occurs. Coarse agglomerate, dipping 10° to 20° N.W. outcrops along the west side of Hillsborough Bay and is well exposed in a bluff at the mouth of the Hillsborough River. Eastward from the river, coarse agglomerate was not noted in the Bacolet formation, except for a small patch forming the tip of Granby Point.

Locally, the agglomerates and tuff-breccias are characterized by the presence of blocks and lapilli of bright- to dark-red feldspar porphyry, colored by finely divided hematite scattered through a microcrystalline groundmass. Concentrations of the red porphyry fragments occur in two belts, one between the tip of Bacolet Point and the west side of Hillsborough Bay, the other extending north and east from Red Point (2 miles west of Scarborough) through Lambeau village and Government Farm to an exposure at about the 1-mile marker on Orange Hill road. Presumably, the two belts are made up of the debris from two or more distinct volcanic eruptions, but whether from the same or different centers was not determined.

Tuffs and fine tuff-breccias cover a far larger area than do the coarser ejectamenta. In general, however, the tuffs are soft, jointed, and deeply weathered, so that outcrops are few and difficult to interpret. In unweathered exposures, as in the sea cliffs of the South Coast, the tuffs are fine, granular, yellowish-green rocks, with more or less abundant crystals and fragments of black pyroxene and greenish feldspar, and with lapilli of feldspar and pyroxene porphyry. Most exposures show intricate fracturing and small-scale faulting, with veinlets of quartz, calcite, prehnite, and zeolite (laumontite or stilbite) sealing the fractures. Irregular veins of yellowish-green epidote-quartz rock are common, generally appearing to be emplaced along shear zones. Water-laid tuffs and interbedded water-deposited sediments seem to be absent.

FIGURE 1. DIORITE CUTTING CONTACT-METAMORPHOSED TUFF-BRECCIA
Goldsborough formation, Mangrove Bay.

FIGURE 2. AGGLOMERATE
Bacolet formation, east side Minister Point.

FIGURE 3. XENOLITH OF DIORITE IN TUFF BRECCIA
Merchiston formation, Goat Island

FIGURE 1. DACITE FLOW, ROCKY POINT

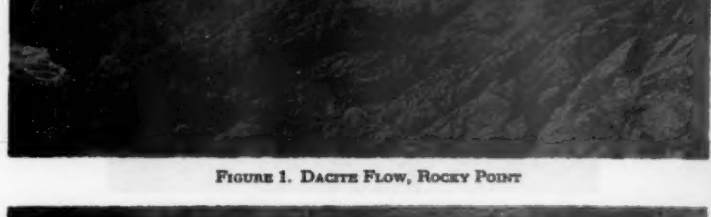

FIGURE 2. PSEUDO-PILLOW STRUCTURE
In dacitic lava, Hawk's Bill.

FIGURE 3. PSEUDO-PILLOW STRUCTURE
View in opposite direction from Figure 2, along gently dipping joint surfaces.

Locally, welded tuffs and tuff-breccias form prominent outcrops. The inlier of volcanics protruding above the coral at Mount Pleasant is a mass of welded andesitic crystal tuff, and a similar resistant mass of tuff-breccia forms a conical hill one-half mile southeast of Plymouth. The waterfall in the Hillsborough East River, 2 miles north of Mount St. George, probably is due to a bed of welded tuff. Less conspicuous occurrences are found among the softer tuffs along the Bacolet and Mount St. George coasts, and in road cuts in the Orange Hill area.

At a number of places, vertical bodies of agglomerate cut bedded pyroclastic rocks of the Bacolet formation. On Minister Point, for example, three such bodies occur, each about 50 feet thick. However these probably are only two intrusions (Fig. 2) and faulting has duplicated one of them. The agglomerate, consisting of blocks of black and red porphyry in a tuff matrix, is coarse and well indurated, and forms pier-like masses extending 50 feet or more from the sea cliff. The presence of blocks of red porphyry suggests that these agglomerate bodies may have been injected at the time of the extrusion of the near-by red porphyry-bearing agglomerates and tuff-breccias and may, in fact, fill the vents from which these pyroclastics were derived.

About three quarters of a mile southeast of Scarborough another vertical agglomerate mass appears in the sea cliff (Fig. 2). Here hard, coarse pyroxene porphyry agglomerate (lacking red porphyry blocks) forms a point of land and extends up the hill as a pronounced ridge to a point 200 yards southeast of old Fort George. The arrangement of dikes and faults suggests this was a center of volcanic activity though not the one from which the adjacent pyroclastics were ejected.

Three other elliptical masses of agglomerate, apparently representing vertical vent agglomerates, were mapped in the Bacolet area (Pl. 1). One is found a quarter mile southeast of Fort George, another just north of Bacolet House, and the third about 2 miles east of Scarborough on Windward road. All have their long axes striking northeast. Although structural relations indicate that these vent fillings near Bacolet are of about the same age, they are younger than those of the Minister Point area (Fig. 2).

Another body of vent agglomerate was mapped half a mile east of Bethel on the Orange Hill road, where it forms an elliptical hill covered with masses of coarse, greenish-black, pyroxene porphyry agglomerate similar to that in the Bacolet area. Except for scattered boulders along the Bethel road, this agglomerate mass seems to be entirely surrounded by fine tuffs dipping in a westerly direction. These tuffs very probably came from centers in the Bacolet area; the near-by agglomerate boulders are best explained as erosional remnants of the pyroclastic debris extruded through the Bethel vent onto the surface of the older tuffs.

Two pipelike bodies of intrusive breccia were observed on the east side of Bacolet Point (Fig. 2). Each is about 20 feet wide, something over 40 feet long, and is intruded into less resistant tuff, making prominent ridges in the sea cliff. The breccias consist of angular blocks up to 3 or 4 inches long in a glassy, tuffaceous matrix. The blocks are green and black glass with phenocrysts of diopsidic augite to 5 mm. long and with somewhat smaller, badly altered plagioclase phenocrysts. Augite crystals reaching a length of 15 mm. occur in the enclosing tuff. Lacroix (1904, p. 156) described similar "breches ignee" injected into crevasses in Nuees ardentes deposits of Mt. Pelee during the eruptions of the winter of 1902–1903.

Lava flows are a minor constituent of the Bacolet formation. In fact, flows were identified with certainty at only one locality, Hill 222 north of Mount Pleasant, where they form the top of the hill. The base of the lava is not exposed. However, tuffs outcrop on the north side of the hill at an elevation of about 150 feet, while lava occurs as low as 180 feet, indicating for it a thickness of 40 to 70 feet.

At the lower elevation the flow is represented by a dense black basalt with reddish-brown patches and streaks, and with phenocrysts of feldspar (labradorite) and diopsidic augite to 2 mm. long. Near the top of the hill the basalt becomes highly vesicular, and has abundant phenocrysts of feldspar (bytownite) about the same size as those in the lower exposure, but pyroxene phenocrysts are lacking. The source of the lava is not apparent. It may have come from the vent near Bethel, now plugged with basaltic agglomerate, but proximity is the only evidence for such a conclusion.

The Bacolet formation includes pyroxene andesite, basalt, olivine basalt, and their glassy and tuffaceous equivalents. Of these, andesite is by far the most abundant, while olivine basalt is quite rare.

In the andesites, microphenocrysts and microlites of andesine and calcic oligoclase are abundant, the rocks tending to have a microcrystalline texture. Coarser plagioclase phenocrysts (andesine or labradorite) have been extensively altered to epidote or carbonate, suggesting that the unaltered feldspar was much more calcic. The few basalts studied, on the other hand, are glassy rocks with comparatively unaltered labradoritic or bytownite phenocrysts. Microphenocrysts are not common, though the glass may be crowded with feldspar microlites. In general, the rocks of this formation lie close to the andesite-basalt boundary and for many specimens a chemical analysis would be necessary to determine the correct classification.

A monoclinic pyroxene having uniform optical properties characterizes all of the Bacolet volcanics (Table 4). The mineral is grayish to colorless, with $Z \wedge c = 40°$ to 42°, $(+)2V = 57°$ to 60°, $\alpha = 1.677–1.681$, $\gamma = 1.705–1.711$, and is diopsidic augite, close to salite (Hess, 1941), a common pyroxene in effusive rocks. Some of the crystals have iron-rich margins, with extinction angles as high as 45°. For the most part the pyroxene is fresh, though a few specimens show partial alteration to yellowish-green serpentine. Hornblende, either as a primary constituent or as an alteration of pyroxene, is uncommon except near the diorite batholith where the hornblende may be explained as a contact metamorphic effect. Neither orthorhombic pyroxene nor pigeonite were identified in any of the thin sections.

Magnetite and ilmenite occur as crystals or disseminated dust in all of the sections except the red porphyry, where the iron oxide is in the form of finely divided hematite. Quartz is sparingly present as minute grains in the groundmass of some of the specimens. Chlorite, carbonate, epidote, and zeolites are common alteration products, and in most of the rocks amygdules or microamygdules are present, filled with quartz, chlorite, or carbonate, less commonly with zeolite or epidote.

Tuffs of the Bacolet formation vary considerably in appearance, depending on the coarseness and composition of the constituent particles. Lapilli-tuffs, crystal-vitric, and vitric-crystal tuffs are the most common varieties. All, however, are of andesitic composition.

Fragments of partially reconstituted tuff, closely resembling normal tuff in hand

specimen, are fairly common in the agglomerates and tuff-breccias. Microscopic examination reveals a hornfelsic groundmass of tiny monoclinic pyroxene and plagioclase crystals, with magnetite, brownish-green hornblende, sphene, serpentine, and zeolite as minor constituents. The pyroxene and feldspar phenocrysts both show zoning, and in many of the feldspars a corroded, cloudy core of andesine or labradorite is present, surrounded by clear, zoned oligoclase-andesine. More thoroughly recrystallized specimens, resembling diorite, have a coarser groundmass, with prominent crystals of hornblende. The welded tuff-breccia forming the conical hill one-half mile southeast of Plymouth contains numerous fragments of such reconstituted tuffs.

The hornfelsic texture of these reconstituted tuffs suggests that they were subjected to high temperatures for a time long enough to permit recrystallization of the tuffaceous material, essentially in the solid state. Presumably, this occurred within or very close to the volcanic throat, where the rocks would be maintained at an elevated temperature for considerable periods of time. Macdonald (1944) has described an analogous occurrence of picritic basalt with hornfelsic texture from the throat of Kilauea volcano, where evidence indicates the parent material had remained for some time in an incandescent state.

*Basalt Tb 3.* The sample, a block of greenish-black pyroxene porphyry, was collected from coarse agglomerate exposed in the sea cliff below Bacolet House. In the hand specimen phenocrysts of pyroxene to 5 mm. in length and smaller phenocrysts of feldspar are scattered through an aphanitic groundmass. The thin section shows euhedral phenocrysts and broken fragments of colorless pyroxene (salite, *see* Table 4), and small phenocrysts and microphenocrysts of plagioclase in a microcrystalline groundmass. The larger feldspar phenocrysts, strongly zoned and somewhat altered, vary from $An_{70}$ at the core to $An_{44}$ at the margin, while one of the larger zoned microphenocrysts has a core of composition $An_{48}$ and a much more sodic margin. A zoned crystal with a badly altered center contains corroded fragments of albite ($An_8$) in the core with successive outward zones of $An_{70}$ and $An_{44}$. (The albite is probably a xenocryst, but whether or not derived from a possible schist basement could not be determined. No xenoliths of schist were found in the volcanics.) Tiny feldspar laths, with possibly a little quartz, form the groundmass; epidote, carbonate, and magnetite are minor constituents. Chlorite-filled micro-amygdules are abundant in parts of the section. The analysis and norm for this sample are given in Table 3.

*Olivine Basalt Tb 15.* A block of dark-red aphanite, collected from medium coarse agglomerate at Bacolet Point, contains phenocrysts of olivine to 2 mm. long and smaller phenocrysts of greenish feldspar. The thin section reveals the presence of euhedral olivine crystals, associated with fairly fresh plagioclase phenocrysts and a very few pyroxene phenocrysts in a cryptocrystalline groundmass of devitrified glass. The olivine has been altered to green antigorite and a very pale green nearly isotropic mineral, possibly penninite. Veinlets of orange-red iron oxide cross the altered crystals and form an opaque to translucent margin around them. A fresh pyroxene inclusion was noted in one altered olivine crystal. Feldspar phenocrysts are bytownite, and the pyroxene is the colorless diopsidic variety (Table 4). The groundmass contains irregular micro-amygdaloids of calcite and is colored by translucent orange-red dust, presumably hematite.

The block of olivine basalt is associated with blocks of basaltic, greenish-black pyroxene porphyry. Both rest in a matrix of dark-green pyroxene prophyry, which under the microscope is seen to be a welded vitric-crystal tuff. The tuff is made up of devitrified, dirty yellow glass, crowded with microlites of feldspar and pyroxene, and with micro-amygdaloids of carbonate or quartz and chlorite. Magnetite, but no finely divided hematite, is present in the glass. Pyroxene and plagioclase occur as abundant euhedral crystals and fragments in the glass matrix. These two minerals are nearly identical with those of the olivine basalt (Table 4).

*Pyroxene Andesite Tb 196.* This sample is a block of greenish-gray, aphanitic, amygdaloidal rock from coarse agglomerate exposed at the mouth of Hillsborough West river. Microscopically the rock has a very fine-grained holocrystalline, intergranular texture, with monoclinic pyroxene, epidote, magnetite, leucoxene, and a very little quartz in a network of minute plagioclase laths. Scattered microphenocrysts of zoned pyroxene and more or less completely altered plagioclase are present.

TABLE 3.—*Chemical analysis and norm of basaltic pyroxene porphyry*

R. B. Ellestad, Analyst

| Chemical Analysis | | Norm | | |
|---|---|---|---|---|
| | | Weight per cent | | |
| $SiO_2$ | 49.80 | Qtz | 2.82 | |
| $Al_2O_3$ | 12.96 | Or | 2.78 | $Or_4Ab_{20}An_{76}$ |
| $Fe_2O_3$ | 3.61 | Ab | 13.62 | |
| FeO | 6.30 | An | 27.80 | |
| MgO | 8.94 | Di | 28.98 | $Wo_{30.8}En_{54.4}Fs_{14.8}$ |
| CaO | 12.93 | Hy | 17.29 | |
| $Na_2O$ | 1.55 | Ap | 0.34 | |
| $K_2O$ | 0.44 | Ilm | 1.22 | |
| $H_2O^+$ | 2.49 | Mt | 5.34 | |
| $H_2O^-$ | 0.24 | Cal | 0.10 | |
| $CO_2$ | 0.03 | | | |
| $TiO_2$ | 0.63 | | | |
| $P_2O_5$ | 0.11 | | | |
| MnO | 0.19 | | | |
| Total | 100.22 | Total | 100.29 | |

The pyroxene is colorless diopsidic augite in the center, becoming more iron rich at the margins. Altered feldspar phenocrysts are andesine ($An_{40}$), the microlites somewhat more sodic, probably calcic-oligoclase. The amygdaloids are highly variable; most are of calcite or of calcite with quartz centers. Other varieties noted include quartz with epidote centers, and quartz-zeolite and quartz-chlorite intergrowths.

In the absence of nonvolcanic sediments and of abundant lava flows, the Bacolet formation is believed to have been formed largely by explosive volcanism, the ash and larger fragments of volcanic rock being deposited subaerially on and near the flanks of the parent volcanoes. Although the mud-flow type of deposit was not recognized, some of the more chaotic assemblages of agglomerates and tuff-breccias could have been formed in this manner.

The probable locations of ancient centers of eruption are indicated by the distribution of vent agglomerates, the dip and strike of the pyroclastic deposits, and the relative distribution of fine and coarse pyroclastic rocks. The coarser agglomerates and tuff-breccias occur in the vicinity of Bacolet Point, while the finer deposits cover a large area to the northwest, north, and northeast. In general, the beds in the Bacolet area dip 10° to 45° to the northwest. Evidently these coarse deposits were derived from a source not very far southeast of Bacolet Point. The possibility that the vent agglomerates of Minister Point may be contemporaneous with the extrusive activity responsible for the pyroclastic rocks near Bacolet has already been mentioned.

A cluster of vertical pipes of vent agglomerate cut the northwest-dipping, coarse pyroclastic rocks north and northwest of Bacolet Point (Fig. 2; Pl. 1). These pipes are younger than the strata they cut, and hence are probably younger than the Minister Point vents. If any large amount of agglomerate or coarse tuff-breccia was extruded through the Bacolet vents, it has been removed by erosion or buried by later tuffs. Part of the tuff which covers the inland area could have originated here.

*Merchiston formation.*—Merchiston pyroclastic rocks outcrop on Little Tobago Island and Goat Island, and on the peninsula between Tyrrel's Bay and King's Bay. The formation is named for Merchiston Estate which occupies most of this peninsula. An area of volcanic tuffs lying north of Carapuse Bay, and apparently forming a roof pendant in the diorite batholith, may belong to either the Bacolet or Merchiston formation; since it is closer to the latter, it is arbitrarily included with the Merchiston rocks.

The agglomerates, tuff-breccias, and tuffs assigned to the Merchiston formation are physically and petrographically identical with the Bacolet volcanics. A separate center of extrusion is indicated, however, by the presence of agglomerates in the vicinity of Tyrrel's Bay. The coarsest of the agglomerates are exposed in the sea cliffs on the south side of the Bay. Subangular to rounded blocks of dark-gray and greenish-black pyroxene and feldspar porphyry, lying in a matrix of porous, yellowish- to blackish-green tuff, average about 6 inches in diameter, with the larger ones measuring about 1 foot across. Toward the south, the agglomerate grades into tuff-breccia, and in Lucy Vale Bay the largest blocks measure about 4 inches in diameter, the average size being about half an inch.

Fine tuff-breccia covers most of the Merchiston Peninsula south of Lucy Vale Bay. On the north side of King's Bay, however, there are large boulders of coarse agglomerate at the base of the sea cliff, in a zone intricately intruded by diorite dikes. The exposure is too poor to permit determining whether this is a vent agglomerate, but its isolated occurrence would favor such a conclusion.

Goat Island is made up of weathered lapilli-tuffs intricately interlaced with calcite-quartz veinlets. Near the east end of the island, on the south coast, the tuffs include a xenolith of diorite, the only fragment of nonvolcanic rock found in the Tobago volcanics. The xenolith is an angular block measuring about 7 by 10 inches. (Pl. 2, fig. 3). Grain size is uniform over its exposed surface, and the calcite-quartz veinlets cross into it from the volcanics. In thin section the xenolith appears to be a medium coarse hornblende diorite, composed of cloudy sodic oligoclase and pleochroic green hornblende arranged in a hypautomorphic-granular texture. A few small crystals of magnetite and apatite are also present. The diorite xenolith rests in a massive lithic-crystal tuff composed of fragments of glass and broken crystals of augite, badly altered andesine, magnetite, and hornblende in a matrix of devitrified glass. One hornblende seen in this section is a large broken crystal identical to the hornblende of the diorite, and is probably a xenocryst.

Pyroclastic rocks on Little Tobago include tuff-breccias and fine agglomerates. They differ from those of Goat Island and the Merchiston area in containing lapilli and small blocks of red feldspar porphyry. The content of red porphyry increases, and the rocks become slightly coarser, from west to east across the island. In the

TABLE 4.—*Optical properties of minerals from selected Tobago rocks*

| Sample No. Formation | Rock Type | Plagioclase | Pyroxene | Amphibole | Other Minerals |
|---|---|---|---|---|---|
| *Schist* | | | | | |
| TB 179 Parlatuvier | epidote actinolite schist | none | none | actinolite: X & Y—colorless, Z—pale grn. (−)2V–80° (approx.) Z∧c—15° | |
| TB 243 Parlatuvier | albite-epidote-actinolite schist | albite—indices below balsam (+)2V | none | actinolite: X—colorless, Y—yel. grn., Z—blu. grn. Z∧C—18°, $\alpha$—1.636, $\gamma$—1.656 (±.003) | |
| TB 247 Parlatuvier | albite-clinozoisite schist | albite—Ab 93–94, (+)2V, $\alpha$—1.529, $\gamma$—1.539 (± .001) | none | | clinozoisite: pale yel., (+) 2V–80°, $\alpha$—1.713, $\gamma$—1.726 (± .001) |
| TB 317 Parlatuvier | amphibolitic contact facies | calcic oligoclase or sodic andesine | none | hornblende: X—pale yel., Y—yel. grn., Z—grn. blue, (−)2V–80°, Z∧c—19° | |
| TB 307 Parlatuvier | amphibolitic contact facies | none | none | hornblende: X—pale yel., Y—olive grn., Z—blu. grn., Z∧c—17°, $\alpha$—1.628, $\beta$—1.636, $\gamma$—1.648 (±.002) | |
| *Ultramafics* TB 192 | pyroxenite | none | augite: grayish, nearly colorless, (+)2V–55°, 56°, Z∧c—44° | none | |
| TB 194 | contact-meta. peridotite | none | augite: grayish, nearly colorless, (+)2V–51°, 53°, 54°, Z∧c—44°, $\alpha$—1.678, $\beta$—1.689, $\gamma$—1.708 (±.002) | uralite: X—pale yel., Y—pale grn., Z—pale brn. grn., (−) 2V–83°, Z∧c—26°, $\alpha$—1.641, $\beta$—1.653, $\gamma$—1.659 (±.002) | olivine: (+)2V, near 90°, serpentine: pale yel., moderate (+) 2V, $\beta$—1.581, mod- |
| TB 202 | contact-meta. pyroxenite | none | augite: nearly colorless, (+)2V–58°, 58°, 56°, Z∧c—41° | uralite: X—pale yel., Y—grn. brn., Z—blu. grn., Z∧c—22° | |

| | | | | | |
|---|---|---|---|---|---|
| | dotite | | 44°, α—1.678, β—1.689, γ—1.708 (±.002) | 2V–83°, ZΛc—26°, α—1.641, β—1.653, γ—1.659 (±.002) | yel., moderate (+) 2V, β—1.581, mod- |
| TB 202 | contact-meta. pyroxenite | none | augite: nearly colorless, (+)2V–58°, 58°, 56°, ZΛc—41° | uralite: X—pale yel., Y—grn. brn., Z—blu. grn., ZΛc—22° | |
| TB 205 | peridotite | none | augite: grayish, non-pleochroic, (+)2V–49°, 49°, 51°, 52°, 54°, ZΛc—42°–43°, α—1.677, γ—1.708 (±.002) | | olivine: 2V near 90°, γ—1.715 (±.003) |
| TB 259 | sheared and contact-meta. pyroxenite | albite (secondary) | augite: dirty gray, non-pleochroic, (+)2V–54°, 56°, ZΛc—43° | none | |
| TB 309 | sheared and contact-meta. serpentinite | none | none | actinolite: X—pale yel., Y—brn. grn., Z—pale blu. grn., ZΛc—20°, 2V near 90° | epidote veinlets, and possibly some chlorite are present. |
| *Volcanics* | | | | | |
| TB 3 Bacolet | basalt block from agglomerate | phenocrysts: bytownite to andesine, avg. labradorite | diopsidic augite, near salite. (+)2V–60°, ZΛc—42°, α—1.677, β—1.690, γ—1.705 (±.002) | none | |
| TB 15 Bacolet | red basalt block from agglomerate | microphenocrysts: An 81–83, α—1.567, γ—1.578 | augite, colorless, not abundant, (+)2V–58° | none | serpentine pseudomorphs after olivine in glassy ground mass. |
| | black basalt block, same hand specimen as red. | phenocrysts: An 79–82 α—1.567, γ—1.577 (±.001) | diopsidic augite, nearly colorless, (+)2V–58°, α—1.681, γ—1.711 (±.001) | none | glassy ground mass, no olivine |
| TB 61 Hawks Bill | dacite flow | phenocrysts: An 26–37 by albite-carlsbad method | none | none | glassy ground mass, abundant skeletal magnetite |

TABLE 4—*Continued*

| Sample No. Formation | Rock Type | Plagioclase | Pyroxene | Amphibole | Other Minerals |
|---|---|---|---|---|---|
| *Volcanics*—Cont'd<br>TB 69<br>Bacolet | basaltic vent agglomerate | microphenocrysts: An 82–85, microlites: about An 55, by albite-carlsbad method | diopsidic augite: pale yel. grn., faintly pleochroic, (+)2V–58°, Z∧c—41° | none | groundmass: vesicular brown glass |
| TB 196<br>Bacolet | pyroxene andesite block from agglomerate | microphenocrysts: about An 40, microlites more sodic, about calcic oligoclase | diopsidic augite: pale yel., nearly colorless, zoned; center of crystal, (+)2V—57°, Z∧c—40°; edge of crystal, (+)2V–55°, Z∧c—43° | none | |
| TB 93<br>Bacolet | contact-meta. crystal-vitric tuff | microphenocrysts, prob. recrystallized, An 35 at center, margins slightly more sodic | augite: yel. gray, faintly pleochroic, (+)2V–52°, Z∧c—41°, $\alpha'$—1.69, $\gamma'$—1.71 | uralite, and secondary hornblende in hornfelsic grd. mass, X—pale yel., Y—olive grn., Z—grn. blue, Z∧c—18°, large (−)2V | |
| *Diorite*<br>TB 94 | augite meladiorite | An 44 by indices | augite: pale yel. gray, non-pleochroic, (+)2V—55°, 57°, Z∧c—41°, $\gamma'$—1.71 | hornblende: X—pale yel., Y—grn. brn., Z—brn. grn., Z∧c—23°, (−)2V large. | |
| TB 100 | meladiorite | slightly zoned crystal, center An 46 by albite-carlsbad method | none | hornblende: X—pale yel. brn., Y—grn. brn., Z—clear grn., Z∧c—26°, (−)2V about 80° | hornblende partly replaced by biotite-epidote intergrowth |
| TB 157 | hornblende gabbro | slightly zoned crystals, An 64–67 by albite-carlsbad method | augite: unaltered remnants, (+) 2V–55°, Z∧c—42–43° | hornblende: brownish green, pleochroic both uralite and primary hornblende present | |
| TB 80 | mesocratic hornblende diorite | zoned crystal, An 63 at center, An 14 at edge; | none | hornblende: X—pale yel., Y—brn. grn., Z—clear green, Z∧c | |

| | | | | | |
|---|---|---|---|---|---|
| | | 64–67 by albite-carlsbad method | 2V–55°, Z∧c—42–43° | ochroic both uralite and primary hornblende present | |
| TB 80 | mesocratic hornblende diorite | zoned crystal, An 63 at center, An 14 at edge; unzoned crystal, An 31 | none | hornblende: X—pale yel., Y—brn. grn., Z—clear green, Z∧c—19°, 2V near 90° | |
| *Dike rocks* | | | | | |
| TB 7 | basalt | microphenocrysts: An 74–77 by indices; crystallites, An 62 by ext. angle | augite: pale yel. gray, non-pleochroic, (+)2V—56°, Z∧c—43°, $\alpha$—1.681, $\beta$—1.690, $\gamma$—1.711 (±.001), p>v, weak | none | |
| TB 17 | augite andesite | phenocrysts and microphenocrysts calcic oligoclase (badly altered) | augite: nearly colorless, (+)2V–52°–55°, Z∧c—42°–44°, $\alpha$—1.685, $\gamma$—1.713 (±.003) | none | |
| TB 38 & 39 | augite-quartz diorite | phenocrysts, andesine (badly altered) | augite: crystals in grd. mass only, (+)2V–54°, Z∧c—44° | hornblende microphenocrysts, X—pale yel., Y—grn. brn., Z—grn. tan, Z∧c—20° | quartz grains and serpentine pseudomorphs after olivine abundant |
| TB 147 | labradorite andesite | zoned crystals of grd. mass, An 75 to An 55, avg. about An 65–70 | none | hornblende: euhedral crystals, brn.-grn., pleochroic | |

sea cliffs surrounding the bay on the east side, red porphyry is particularly abundant, forming about a quarter of the agglomerate blocks, the remainder being dark-green and gray porphyries. The blocks are angular and of fairly uniform size, average about 2 inches long, and lie in a lithic tuff matrix.

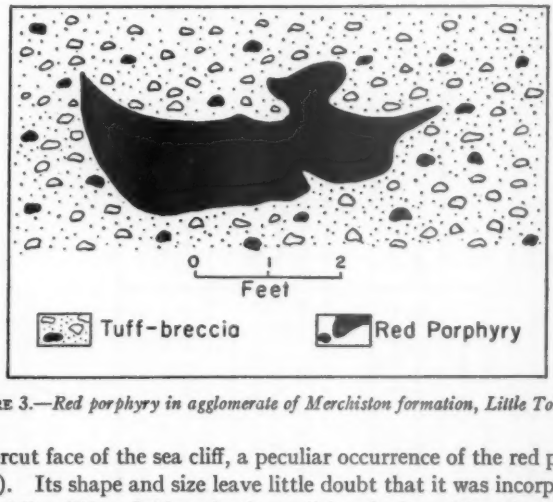


FIGURE 3.—*Red porphyry in agglomerate of Merchiston formation, Little Tobago*

In an undercut face of the sea cliff, a peculiar occurrence of the red porphyry was noted (Fig. 3). Its shape and size leave little doubt that it was incorporated in the agglomerate while still in a fluid or plastic condition, for it could not have been ejected explosively or otherwise mechanically incorporated in the enclosing agglomerate. Most likely the agglomerate here lies in the throat of an ancient volcano and the red porphyry represents either new or re-fused lava which was never ejected.

Volcanics forming the roof pendant in the diorite north of Carapuse Bay consist of light to dark-gray tuffs and fine tuff-breccias with fragments of pyroxene and feldspar porphyry. Numerous apophyses from the diorite batholith cut the volcanics, which commonly show slight contact-metamorphism, evidenced by recrystallization of the fine groundmass and development of green hornblende. Contact-metamorphism of the volcanics is discussed more fully in another part of this paper.

The presence of coarse agglomerate on the south coast of Tyrrel's Bay, grading southward into finer pyroclastics, suggests a volcanic vent in the vicinity of the bay. The only indication of a vent noted in this area is on Little Tobago, but the lack of red porphyry in the Tyrrel's Bay rocks makes it doubtful that these pyroclastics came from the Little Tobago vent. The possibility of an occurrence of vent agglomerate on the north side of King's Bay was noted above, but it is unlikely that this is the source of the coarse agglomerate to the north, because the grain-size gradation is in the wrong direction. Apparently the ancient vent lies off the present shore, most likely within or near Tyrrel's Bay.

*Hawk's Bill formation.*—A distinctive group of dacitic lava flows outcrops on the northwest coast of Tobago, south of the town of Plymouth. The northernmost exposure of the flows occurs in a road cut a quarter mile south of Plymouth, where dark-gray, aphanitic, feldspar porphyry, dipping gently eastward, overlies weathered

tuffs resembling those of the Bacolet formation. To the south and east, the limits of the Hawk's Bill formation are indicated by outcrops of Bacolet-type tuffs south of Rocky Point and along the trail leading southeast from the village of Black Rock to the Orange Hill road. Excellent exposures of the flows may be seen on Rocky Point and on Hawk's Bill, which is designated the type area of the formation.

Along the shore road, about half way between Hawk's Bill and Rocky Point, the cuts contain deeply weathered, dark greenish-gray lapilli-tuff. These pyroclastics, extensively faulted and sheared, have been intruded by small, irregular bodies of black, very coarse feldspar basalt porphyry, which is also found elsewhere cutting the Bacolet volcanics. Whether these tuffs should be included in the Hawk's Bill or Bacolet groups is uncertain. Their petrographic character, and the fact that they lie beneath the flows, favors the latter correlation.

On Rocky Point is exposed a series of flows of dark gray porphyritic dacite, blocky and scoriaceous in part, in part exhibiting a laminar structure (Pl. 3, fig. 1). A bright red soil results from the weathering of these rocks, contrasting sharply with the chocolate brown soils derived from the other Tobago volcanics.

Just north of Rocky Point irregular lenticles of red jasper are found in the flows, which are here brownish-green porphyries with small phenocrysts of brownish feldspar. Thin sections of these rocks show that they are devitrified perlitic glass. The glassy character of the rock, indicating rapid chilling, together with presence of jasper, suggests subaqueous extrusion. Unfortunately, the jasper-bearing outcrop is very small and pillow structure or other confirmatory evidence of under-water consolidation was not observed.

Dark-gray porphyritic flows outcrop on Hawk's Bill and along the coast to the northward. On the south side of Hawk's Bill a small tongue of porphyritic dacite exhibits a peculiar type of structure superficially resembling pillow-lava (Pl. 3, fig. 2). The "pillows" consist of fine-grained, dark gray feldspar porphyry, showing no evidence of concentric or radial structures, or of chilled margins. They appear to be elongated blocks (Pl. 3, fig. 3) included between joints of two systems, one nearly vertical and at right angles to the elongation of the lava tongue, the other nearly horizontal but dipping gently eastward through the tongue. Extensive alteration along these joints has produced an epidotic margin around the dacite blocks (Pl. 4, fig. 1).

Dark-gray, fine-grained porphyritic dacite predominates in the Hawk's Bill formation. The white feldspar phenocrysts average about 1 mm. long. Parts of the flows have amygdaloids of crystalline quartz, some with hollow centers showing terminated quartz crystals. In thin section the dacites are seen to have a fine-grained to cryptocrystalline ground mass of interlocking feldspar and quartz grains with abundant skeletal crystals of magnetite and a little chlorite. The feldspar microlites tend to be oriented in a crude flow structure. Hornblende may be present as small, partly resorbed, brownish-green prisms. Feldspar occurs as phenocrysts and glomeroporphyritic aggregates of sodic andesine ($An_{27-37}$), rather fresh in appearance but with somewhat corroded margins. Feldspars of the ground mass have indices below quartz and are therefore sodic oligoclase or albite. Two types of amygdaloids are present, the larger being quartz, the smaller, green chlorite with

rims of finely granular quartz. Abundant finely divided magnetite gives the fresh rock its gray color.

The epidotized margins of "pseudo-pillows" (south side of Hawk's Bill) have textures much like that of the adjacent unaltered dacite, but the groundmass has been changed to finely granular epidote and quartz, with some albite. Magnetite has disappeared and the feldspar phenocrysts are badly corroded. Apparently, magnetite reacted with the anorthite molecules of the feldspar to form epidote, very likely under the influence of volatiles given off by the flow as it cooled.

The jasper-bearing rock from the north side of Rocky Point (Tb 64), is unusual in that it has been recrystallized without destroying its perlitic texture. In plain light the thin section shows a few feldspar laths scattered through a dirty yellowish-brown, perlitic, and apparently glassy groundmass (Pl. 4, fig. 2). Under crossed nicols, however, the rock is seen to be holocrystalline, with the perlitic texture preserved by the crystallization of dirty, brownish quartz in layers concentric to the perlitic cracks. Epidote and chlorite may be present in layers concentric to the quartz, and granular epidote fills the centers of many of the concentric structures (Pl. 4, fig. 3).

The Hawk's Bill flows overlie tuffs of the Bacolet formation and seem to be definitely younger than, rather than interbedded with the tuffs. The tuffs commonly are deeply weathered and strongly sheared, with quartz-epidote and quartz-carbonate veins developed along shear and joint planes; the flows are fresh, unsheared, and show only minor faulting, as though they had escaped much of the stress that affected the other Tobago volcanic rocks.

The prevailing westerly dips of the flows seem to indicate a source somewhere east of Hawk's Bill. There are numberous andesite and dacite dikes to the east and southeast of Plymouth, suggesting a possible center of extrusion in this area. The dikes cut the diorite batholith, which has intruded and metamorphosed the Bacolet, Goldsborough, and Merchiston volcanics. If the Hawk's Bill flows are related to the andesite dike system, they are the youngest extrusive rocks of Tobago.

### DIORITE BATHOLITH

*General statement.*—The diorite batholith of Tobago outcrops in a nearly continuous belt 1 to 3½ miles wide, extending from the northwest coast at Plymouth to the southeast coast at King's Bay, and is interrupted only by an area of Merchiston volcanics near Roxborough. These volcanics are intricately intruded by the diorite and are believed to be a roof pendant above the batholith. The presence of diorite exposures in the vicinity of Mangrove Bay indicates the batholith may also underlie a portion of the Bacolet and Goldsborough volcanics.

*Age relations.*—The diorite is intrusive into the North Coast schist, the ultramafics, and the Goldsborough, Bacolet, and Merchiston volcanics. The nature of these contacts will be discussed in detail later. No contact was observed between the Hawk's Bill volcanics and the diorite body; hence their relative ages are a matter of conjecture. As pointed out in the discussion of the Hawk's Bill formation, however, the volcanics are probably younger.

*Description.*—The diorite is strongly sheared and nearly everywhere deeply weathered and dissected; unweathered exposures occur only along the coast and in deep

FIGURE 1. EPIDOTIC ALTERATION AROUND PSEUDO-PILLOW, HAWK'S BILL.

FIGURE 2. RECRYSTALLIZED PERLITE
North side of Rocky Point. Plane polarized light; q—quartz; p—plagioclase; c—chlorite; e—epidote. x 10.

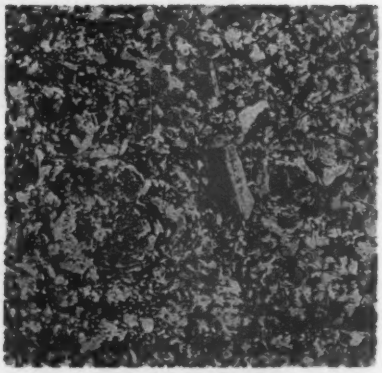

FIGURE 3. SAME AS FIGURE 2.
With crossed nicols. x 10.

HAWK'S BILL FORMATION

FIGURE 1. HOOD OF DIORITE BATHOLITH
Meladiorite brecciated and intruded by lighter-colored diorite.
East side Arnos Vale Bay.

FIGURE 2. HORNBLENDITE IN MESOCRATIC HORNBLENDE DIORITE,
CULLODEN BAY.

FEATURES OF THE DIORITE

artificial cuts. Much of it weathers to saprolite, which retains the texture and color of the parent rock so that mapping is possible even though no unaltered rock remains.

Perhaps the outstanding characteristic of the batholith is its extreme variability. Medium-grained hornblende and pyroxene meladiorite and melagabbro form its margins and also occur in the interior. These dark rocks have been intricately intruded by lighter-colored hornblende diorite, biotite quartz diorite, and a variety of pegmatites and aplites. In the interior of the batholith, outcrops of meladiorite tend to be localized in topographically high areas, with lighter-colored diorite occurring near by at lower elevations. Apparently the dark rocks represent the first phase of the intrusion, the hood of the batholith, which was further intruded and brecciated by less mafic magma (Pl. 5, fig. 1).

Mesocratic hornblende diorite, the second intrusive phase, occurs most commonly inward from the margins of the batholith, though locally it cuts through the meladiorite of the margins and forms the contact rock. Two facies of the mesocratic diorite can be recognized, one an even-grained rock made up of stubby prisms of hornblende and feldspar, the other composed of a mesh of acicular hornblende crystals with interstitial feldspars laths. The latter type forms dikes in the meladiorite. In the Culloden Bay area, on the other hand, the mesocratic facies contains dike-like bodies of very dark, hornblende-rich diorite and irregular stringers of hornblendite (Pl. 5, fig. 2). Both may be pegmatitic phenomena, for coarse, dark, hornblende-rich pegmatite is abundant near the margins of the batholith.

The youngest phase of the intrusive, biotite quartz diorite, outcrops near the center of the batholith as an elongate, irregular body extending eastward from the coast near Plymouth to a point at least 5 miles inland (Pl. 1, Section D-E). This rock weathers to a distinctive tan sandy soil containing shiny flakes of golden mica. Dikes of the biotite quartz-diorite cut both of the earlier phases. They are most common in the center, but are found sparingly throughout the batholith. Away from the center, these dikes may show chilled margins; more commonly they are of uniform texture throughout.

Distinctly gradational contacts between the biotite quartz diorite and mesocratic hornblende diorite were observed in several places, chiefly in the interior of the batholith. The gradation occurs in a very few feet, quartz diorite becoming hornblendic and passing into hornblende diorite of similar texture. Such contacts may be seen on the coast between Plymouth and Arnos Vale Bay, and near the village of Mount Thomas on the road down to Culloden Bay. The fact that its contacts tend to be gradational near the center of the batholith while showing chill features near the margin suggests that the biotite quartz diorite was injected while the center of the batholith was still hot and fluid, thus allowing mixing at the contacts, though the margins of the batholith had cooled and solidified.

Cognate xenoliths of earlier, more mafic diorite are common in the later phases, and xenoliths of schist, volcanics, and ultramafics abound near the contacts of the diorite with these respective rocks. In addition, large xenoliths of volcanics were found north and east of King Peter's Bay, near the schist and $3\frac{1}{2}$ miles north of the nearest known outcrop of volcanics; others occur in the biotite quartz diorite near Plymouth. Schist xenoliths were observed in meladiorite on the west side of King's Bay, and roof pendants of contact metamorphosed schist occur in the diorite 2 miles

to the north. Pyroxenite xenoliths, evidently derived from the ultramafic series, were collected from the diorite just west of Mason Hall, 2 miles southwest of the nearest known ultramafic outcrop. Apparently, therefore, the diorite was intruded through a basement composed of the schist, volcanics, and ultramafics that outcrop elsewhere on Tobago. Two xenoliths of augite-hypersthene gabbro, one collected a half mile southwest and one 1½ miles northeast of Les Coteaux, represent a rock type not otherwise known on the island.

*Petrography.*—Hornblende and augite meladiorites make up most of the mafic marginal facies of the diorite batholith. Hornblende gabbro is also present, though apparently not abundant. In general, these rocks are medium to coarse grained, dark gray to nearly black, and have a hypautomorphic-granular texture. Hornblende makes up as much as 75% of the mafic minerals. Augite is common only in the contact zone of the batholith and invariably shows corroded margins and uralitization. The predominant feldspar is calcic andesine ($An_{42-46}$), complexly twinned, but only weakly zoned. A few sections also have a little clear, interstitial albite and quartz occurring as a product of late crystallization. Biotite, intergrown with clinozoisite or epidote, may be present as a replacement of hornblende. Magnetite and ilmenite make up as much as 5% of these rocks, and apatite and sphene are common accessories.

Although the augite-bearing diorites are commonest in the xenolith-rich zone near the contact with the volcanics, the augite is believed to be a product of primary crystallization of diorite rather than mechanical incorporation of volcanic material. The augite of the diorite differs from that of the volcanics in having smaller optic angle (55° vs. 57°–61°). Perhaps the best evidence is that diorite intruded into schist also carries corroded augite though the schist contains no pyroxene.

The presence of augite in rocks normally hornblendic seems to be fairly common, expecially in contact zones and small intrusive bodies. Buddington and Callaghan (1936, p. 438) cite examples from the dioritic intrusions of the Cascade Range in Oregon. They explain the formation of pyroxene, rather than hornblende and biotite, as being due to rapid cooling and easy escape of volatiles into porous tuffs and flows at shallow depths. Such an explanation would seem to apply to the Tobago occurrence. A further cause of rapid cooling is found in the abundant amphibolitic inclusions, in that the heat and volatiles needed to convert the volcanic tuffs to amphibolites came from the diorite magma and thus hastened its solidification.

*Pyroxene Diorite Tb 94.* This rock outcrops in the bed of Courland River, 1½ miles north of Providence, on the road to Les Coteaux. Contact metamorphosed volcanics occur in road cuts a few yards to the south. The sample is a dark-gray, medium-grained rock carrying inclusions of amphibolite which are probably xenoliths of reconstituted volcanic tuff. A Rosiwal analysis of the pyroxene diorite gives the following mineral composition:

| | |
|---|---|
| Augite | 10.8 |
| Uralite | 39.5 |
| Ilmenite-magnetite | 4.6 |
| Biotite plus Clinozoisite | 2.1 |
| Plagioclase | 42.5 |
| Apatite | 0.5 |
| | 100.0 |

The augite (Table 4) has been largely altered to pleochroic, brownish-green hornblende, none of which appears to be primary. The hornblende, in turn, is partly replaced by an intergrowth of biotite and clinozoisite in which the latter occurs as layers alternating with basal plates of biotite. The plagioclase, calcic andesine, is complexly twinned but only slightly zoned. "Ore minerals" seem to be of two generations. The larger grains crystallized early and were included poikilitically in augite and to a smaller extent in feldspar. A second generation is represented by a graphic intergrowth of "ore minerals" with pale bluish-green amphibole and colorless clinozoisite. This intergrowth seems to have developed in some cases from the alteration of augite. Where augite was in contact with the early generation of magnetite, however, a corona of dark brownish-green uralite formed around many of the magnetite grains. Apatite crystals, representing the earliest stage of crystallization, are included poikilitically in augite, feldspar, and ore minerals.

*Hornblende Meladiorite Tb 100.* The sample was collected from a road cut 100 yards north of Tb 94. It is a massive, medium-grained rock, lighter in color than Tb 94, though still melanocratic. Xenoliths are not common in this exposure. A Rosiwal analysis gives the following mineral composition:

| | |
|---|---|
| Hornblende | 47.4 |
| Magnetite-ilmenite | 1.3 |
| Biotite (plus a little chlorite) | 1.0 |
| Epidote | 5.8 |
| Plagioclase | 44.5 |
| | 100.0 |

A very few grains of apatite and sphene are also present.

Hornblende and feldspar apparently crystallized simultaneously, with a tendency toward euhedrism in the former. However, much of the hornblende occurs as coarse, irregular grains, poikilitically enclosing ore minerals, feldspar, and epidote. Part of the poikilitic hornblende may be uralite, though there is no pyroxene present in the section. Biotite, interleaved with greenish-yellow, pleochroic epidote, has replaced hornblende, and the biotite, in turn, has largely altered to clear green chlorite. The plagioclase, moderately zoned and complexly twinned, is calcic andesine (Table 4). Epidote occurs as primary grains and as an alteration product of feldspar; its association with biotite was described above.

*Hornblende Gabbro Tb 157.* The sample, collected from the west side of Culloden Bay, is a dark-gray, medium-grained rock with abundant fine hornblendic inclusions. These inclusions are probably contact metamorphosed schist, for banded diorite occurring near by closely resembles the diorite-schist migmatite which outcrops on the coast 1½ miles to the east. The rock has a hypautomorphic-granular texture and is made up of brownish-green, pleochroic hornblende, complexly twinned plagioclase and rounded and embayed augite (Table 4). Magnetite-ilmenite and a little apatite are also present. An estimate of the rock's composition gives 50% hornblende, 5% pyroxene, 3% ore and 42% feldspar. The augite is obviously an early mineral, but shows comparatively little uralitisation. Most of the hornblende is primary and crystallized contemporaneously with the feldspar. A small amount of pale bluish-green fibrous amphibole apparently resulted from the deuteric alteration of the hornblende. The slightly zoned plagioclase crystals have centers of calcic labradorite ($An_{64-67}$) and an average composition of medium labradorite.

The equigranular variety of mesocratic hornblende diorite has a hypautomorphic-granular texture resembling that of the meladiorites. As in the latter, the predominant feldspar is a slightly zoned calcic andesine. The brownish-green, pleochroic hornblende seems to be entirely primary and pyroxene is absent. Magnetite, epidote, apatite, and sphene are minor constituents. The other facies, younger than the equigranular type, is characterized by a mesh of acicular hornblende crystals with interstitial feldspar. The hornblende is the typical brownish-green variety, but the feldspars tend to be strongly zoned, ranging in composition from labradorite ($An_{63}$) at the core to sodic oligoclase ($An_{14}$) at the margin, and averaging oligoclase to sodic

andesine. Most samples show up to 10% of micropegmatite, consisting of clear quartz and albite with a little sphene.

The biotite quartz-diorites are leucocratic rocks in which pyroxene is absent and hornblende is present only in zones gradational to hornblende diorite. As in the meladiorite, epidote occurs interleaved with the biotite. Plagioclase shows pronounced zoning, oscillatory in some grains, regular in others, and most are untwinned. Quartz makes up about 25% of the rock. A little magnetite and apatite are present, and one section carries about 2% of pyrite. No orthoclase was observed in any of the sections.

*Biotite Quartz-diorite Tb 149.* The sample was collected from the sea cliff north of Plymouth. It is a medium-grained leucocratic rock with a brownish tinge due to slight weathering of biotite. The mineral composition, in volume per cent, is as follows:

| | |
|---|---|
| Plagioclase | 63.3 |
| Quartz | 26.9 |
| Biotite | 9.0 |
| Epidote | 0.2 |
| Magnetite-ilmenite | 0.6 |
| | 100.0 |

A few grains of apatite are also present.

The rock has a granitic texture. Plagioclase occurs as interlocking, zoned, and twinned grains with oligoclase cores and calcic albite margins. Clear quartz grains are partly interstitial to feldspar and partly contemporaneous with the more sodic outer zones. Biotite, with interleaved yellowish-green epidote, seems to be a relatively late mineral, contemporaneous with the last stages of feldspar crystallization.

Aplite and pegmatite dikes occur throughout the diorite batholith. Pegmatites have been found cutting volcanics and ultramafics near the diorite contact, and are so abundant in the dark, marginal rocks of the batholith that the term "pegmatitic facies" was employed in the field mapping before its marginal character became apparent.

The pegmatites exhibit a variety of colors and textures, varying from white to cream-colored to black, and from medium-grained to very coarse-grained masses having crystals 5 inches or more in length. All are made up of black amphibole and white to cream-colored plagioclase feldspar. Pyrite, epidote, and chlorite are present in small amounts, but no other mineral was identified. The amphibole-rich pegmatites occur, for the most part, as irregular masses a few inches to many feet in greatest dimension, and many of them are cut by dikes of feldspar-rich pegmatite. Some of these amphibolitic masses probably resulted from concentration of hyperfusibles in the nearly-crystallized magma. Others, consisting almost entirely of pargasite, may be xenoliths of completely reconstituted ultra-mafics, though this is only conjecture. The lighter-colored, feldspar-rich pegmatites occur almost invariably as sharply defined dikes, as do the aplites.

The black, prismatic amphibole of the coarse pegmatites seems in every case to be pargasite, with a (+)2V of about 70°, $Z \wedge C = 22°$, X = pale greenish yellow, Y = pale bluish green, Z = clear green. The feldspar, twinned and somewhat saussuritized, varies from medium oligoclase to calcic andesine.

A cream-colored aplite dike half an inch thick, collected from meladiorite at Culloden Bay, contains about 50% feldspar, 40% zoisite, 5% pale-green actinolite, 5% carbonates, and a little apatite and sphene. The feldspar is nearly pure albite. The zoisite, apparently $\beta$-zoisite, has a moderate (+)2V, with the optic plane oriented perpendicular to the cleavage.

*Contact metamorphism.*—Contact metamorphic phenomena attributable to the diorite were observed in the schist, volcanics, and ultramafics. As might be expected, the most pronounced changes occur in the schist, where the soft, light-colored albitic rocks are converted to hard, dark-gray hornfels and amphibolite. Changes in the volcanics are apt to be inconspicuous in the field, but recrystallization of the tuffaceous material to hornfels and uralitization of the pyroxene phenocrysts may be observed under the microscope. In the ultramafics, contact metamorphism resulted mainly in the development of actinolitic hornblende and of epidote veinlets. The diorite-schist contact cuts across the regional strike of the schists so as to bring the igenous rock into juxtaposition with both the Mount Dillon metasediments and the Parlatuvier metavolcanics. The contact-metamorphic facies are somewhat different in these two formations.

In the Mount Dillon formation contact phenomena were observed only on the coast between King Peter's Bay and a point just northeast of Anse Flamengo. Near the contact, diorite exposures exhibit the coarsely pegmatitic autobrecciated facies typical of the margins of the batholith. The contact rocks themselves are gray, finely granular, banded migmatites, some of which contain recognizable stringers of schist.

The more dioritic bands of the migmatite are of medium texture (2 mm ±) and consist of plagioclase and bright-green pleochroic hornblende, with clear yellow crystals of clinozoisite and grains of colorless, slightly birefringent garnet. The plagioclase (andesine) is zoned and twinned as in normal diorite, but the crystals have been fractured and the fragments completely separated and disoriented, suggesting that the feldspar had crystallized and that shattering took place during intrusion of a crystal mush. Hornblende lacks the brownish tinge characteristic of that mineral in normal diorite, and a few rounded fragments of augite are present.

A finely granular aggregate of greenish-brown biotite, green hornblende, saussuritized plagioclase, and magnetite constitute the schistose layers of the migmatite. The biotite and green hornblende seem to have developed as a result of contact action, since neither is present in schist away from the diorite contact.

The nature of the contact indicates that the schistose structure was present at the time of intrusion of the diorite and that the dioritic magma or crystal mush was intruded *lit-par-lit* without altering the orientation of the schistosity to any marked degree. Along the sea cliff between King Peter's Bay and Anse Flamengo, banding in the migmatites strikes slightly north of east, parallel to the predominant strike in the main schist body (Pl. 1). On the north side of Anse Flamengo, the migmatitic banding strikes N. 75 E. and dips 43° S., while 300 feet to the north quartz schist showing no contact effects strikes N. 70 E. and dips 40° S. (Section D-E, Pl. 1). This is approximately the average dip and strike of schistosity within the schist group.

In contrast with the relatively narrow and well-defined contact between the diorite

and the Mount Dillon metasediments, contact effects may be observed over a rather broad area within the Parlatuvier formation, which has been intricately intruded by apophyses from the diorite body. Particularly good exposures occur along the Windward road at Lambeau Hill and along the trails extending northward from the Windward road between Roxborough and King's Bay.

Contact metamorphism of these schists produced a dense hard rock with poorly preserved schistose fabric. Large poikilitic hornblende porphyroblasts developed, as did crystals of garnet and sphene. Albite, if originally present, was converted to a more calcic feldspar (oligoclase-andesine), and this in turn was more or less completely saussuritized. Epidote and clinozoisite appear to have survived the contact-metamorphic processes with little or no change.

Differences between the contact-metamorphic effects in the Mount Dillon and Parlatuvier formations can be ascribed partly to lithologic differences between the two formations. On the other hand, the relatively slight effect of the magma on the Mount Dillon rocks may be due largely to the location of the area on the steep or vertical flank of the batholith well below its roof (Section D-E, Pl. 1); while stronger contact effects noted in the Parlatuvier formation, on Lambeau Hill for example, can be explained by the fact that the batholith underlies this latter area at shallow depth.

Within the volcanics two types of contact metamorphism were noted: alteration (1) to hornblende-andesine hornfels, and (2) to epidote-garnet-sericite rock. By far the commoner type, (1) is associated with the Bacolet and Goldsborough volcanics. Though it may be difficult to recognize in hand specimen, the rock becomes harder and somewhat darker than unmetamorphosed tuff, and on weathered surfaces the white, suassuritized feldspar phenocrysts stand out prominently against the dense, black hornfelsic groundmass. Green quartz-epidote veins are particularly abundant in the tuffs near the diorite contact.

Excellent exposures of the contact facies may be seen along the roads which cross the contact north of Scarborough, especially the Plymouth, Les Coteaux, and Mason Hall roads. In the Les Coteaux road section, unaltered lapilli-tuff outcrops within a quarter of a mile of the contact. A few yards to the north, however, weathered boulders of fine-grained black rock with prominent white phenocrysts appear along the road. The thin section (Tb 92) indicates that this rock prior to metamorphism was a feldspar-rich, crystal-lithic tuff. Contact action by the diorite recrystallized the tuffaceous groundmass to a hornfels of green hornblende and minute, strongly zoned feldspars and caused complete uralitization of the augite. Large feldspar phenocrysts, on the other hand, show little evidence of alteration beyond the development of minute hornblende inclusions. The feldspar is labradorite-bytownite, varying from $An_{87}$ in the core of crystals to $An_{60}$ at the margin.

Further to the north, within 100 yards of the contact, black, dense pyroxene porphyry occurs. This rock (Tb 93) is low in feldspar and apparently represents a metamorphosed pyroxene crystal-vitric tuff. The hornfelsic groundmass, of green hornblende and clear feldspar, is of coarser texture than Tb 92, and the few feldspar micro-phenocrysts present have been almost completely saussuritized. On the other hand, remnants of augite (Table 4) are still present, apparently protected from complete alteration by a thick corona of uralite. The zoned feldspar of the recrystallized

groundmass is oligoclase-andesine, one of the larger crystals having a core of $An_{35}$ and a somewhat more sodic margin.

The actual contact between the diorite and volcanics is covered, but pyroxene meladiorite, carrying abundant hornblendic xenoliths, outcrops in the bed of Courland river. This rock, Tb 94, has been described previously. The xenoliths are similar in appearance to Tb 93, but augite and feldspar phenocrysts have entirely disappeared. An advanced stage of contact metamorphism is shown by a xenolith of pyroxene porphyry collected from biotite quartz diorite a quarter of a mile east of Plymouth on the Les Coteaux road. The rock has been strongly recrystallized so that the pyroxene crystals rest in a dark-gray, medium-grained groundmass of feldspar and hornblende, and the whole rock resembles a meladiorite. In thin section the tuffaceous groundmass is represented by a medium-grained granitoid aggregate of brownish-green hornblende and twinned and strongly zoned plagioclase with a little magnetite-ilmenite. Thick coronas of pleochroic uralite surround grayish augite ($+ 2V = 55°$, $Z \wedge C = 42°$), and the feldspar microphenocrysts have been completely saussuritized. The feldspars of the groundmass show saussuritized cores with nearly clear, probably secondary feldspar around the cores. They vary in composition from labradorite in the center to albite at the margin. An amygdaloid, possibly originally chlorite-carbonate, now has a groundmass of chlorite shot through with needles of pale green actinolite, and contains large grains of zoisite.

The augites of contact metamorphosed volcanics have a somewhat smaller optic angle (52°–55°) that those of the unaltered volcanics (58°–61°), though the angle of extinction remains constant at 41°–43° (table 4). Augite occurring as an early crystallization phase in the diorite has an optic angle of about 55°, near that of the contact rocks. Apparently, contact action tends to change pryoxene of the volcanics into the form stable at the crystallization temperature of the pyroxene-diorite, but the chemical nature of these changes is not apparent from the optical data.

Contact metamorphism of volcanics producing an epidote-garnet-zoisite rock was noted only in the Merchiston formation, the best exposures occurring on Merchiston local road and Windward road at the top of Lambeau Hill. This type of alteration produced a dense, greenish-gray to tan rock, with patches of epidote and pink garnet. Under the microscope the rock has the appearance of a crystal-lithic tuff, the groundmass of which has been altered to a very fine aggregate of zoisite and albite, with scattered specks of magnetite, garnet, and epidote. Feldspar phenocrysts are represented by intergrowths of columnar zoisite crystals. Yellow to yellowish-green epidote and nearly colorless garnet grains up to 2 mm. long are scattered through the groundmass, and veinlets of zoisite are abundant.

### DIKES

Dikes occur more or less abundantly throughout the diorite and volcanics, and a few unmetamorphosed dike rocks were observed in the North Coast schist. They are especially abundant within and near the west end of the diorite batholith, where relatively deep erosion has exposed the central mass of biotite quartz diorite. In the eastern half of the island, on the other hand, where only the hood of the batholith

has been uncovered by erosion, dikes (exclusive of pegmatites and aplites) are extremely rare.

The dike rocks vary from glassy to medium grained, and from labradorite basalts to oligoclase dacites, though andesites are far the most abundant. Most are medium gray to gray green but the aphanitic and glassy varieties may be dense black rocks resembling typical basalts. True basalts are rather rare, however; only one was positively identified, this from the sea cliff on the west side of Bacolet Point.

The dikes have about the same range of composition noted in the diorite and also in the volcanics. Rocks characterized by calcic andesine predominate in all three, quartz-bearing andesine-oligoclase rocks are fairly common, and labradorite or bytownite rocks are present but rare. No lamprophyres were observed.

In general, the coarser dikes resemble normal diorites or quartz diorites, with brownish-green hornblende and oligoclase or andesine feldspar forming the bulk of the rock. Commonly the feldspar has been saussuritized and hornblende partially altered to green chlorite. In extreme cases, the dike may be an oligoclase-epidote-chlorite rock, apparently a result of deuteric alteration, because the surrounding rocks show no appreciable alteration and the dike retains the appearance and texture of an unaltered diorite.

In aphanitic andesite dikes and in the chill zones of coarser dioritic dikes, augite is invariably present and serpentinized olivine pseudomorphs are common. The augite may occur as microphenocrysts in a very fine-grained groundmass of brown hornblende and andesine, or if hornblende is absent, the rock will consist of laths of plagioclase with interstitial augite. The augite is similar to that found in the marginal facies of the diorite. Optical characteristics of minerals from four typical dike rocks are given in Table 4.

### STRUCTURAL HISTORY OF DIORITE AND VOLCANICS

*General statement.*—The diorite and volcanics have undergone a similar structural history, differing on the one hand from that of the older schists and ultramafics, and on the other from that of the younger Tertiary and Quaternary sediments. Two categories of structures can be distinguished, those related to intrusive forces and those developed by stresses of a regional character.

*Structures resulting from intrusive processes.*—Linear and platy flow structures, and primary fracture systems with accompanying dikes and veins, are associated with intrusive igneous masses such as the diorite batholith. However, flow structures are not commonly observed in the Tobago diorite, partly because of its equi-granular texture, partly because of the scarcity of fresh, unsheared outcrops. Linear flow structure in diorite with acicular hornblende crystals was noted a half mile south of Les Coteaux (Pl. 1), the lineation pitching 45° in a northeasterly direction. Flow layering in diorite may be seen on the coast just northeast of Culloden Bay (Pl. 6, fig. 1), and banded migmatites are found in the vicinity of King Peter's Bay.

Inasmuch as primary fractures related to intrusion of the diorite are obscured by post-intrusive deformation, they can now be identified most easily by the dikes emplaced along them. These dikes average about 2 feet and reach a maximum thickness of 10 feet. Most of them strike about east-west and dip toward the axis of the intru-

FIGURE 1. FLOW-LAYERING IN DIORITE
Coast northeast of Culloden Bay

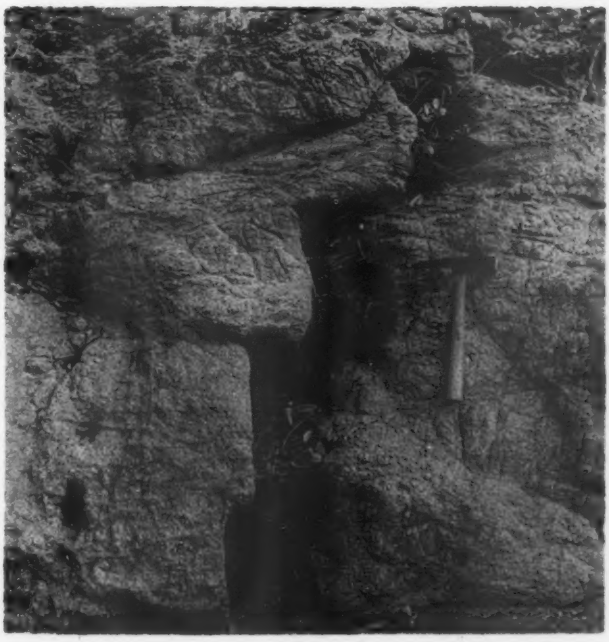

FIGURE 2. FAULTED ANDESITE DIKE IN DIORITE, ARNOS VALE BAY
Coast northeast of Culloden Bay

sion (Pl. 1, Section D-E). Dikes are particularly abundant in the quartz diorite of the center of the batholith, where they have steep to vertical dips and occur in subparallel swarms. (Only the average dip and strike is indicated on Plate 1.) Toward the margins of the batholith the dikes are less abundant and tend to dip more gently with increasing distance from the center. They appear to be arranged in a vertical fan-shaped pattern having its long axis parallel to the elongation of the batholith. A similar fan of tension fractures, with or without accompanying dikes, is commonly present in igneous bodies of batholithic dimensions and is attributed by Balk (1937) and others to continued intrusive pressure after much of the batholith had solidified. According to Balk (p. 99), the tension joints rarely terminate at the contact of any large igneous intrusion, but penetrate portions of the wall rock. In the Sierra Nevada batholith, for example, the fan of cross (tension) joints embraces a strip of wall rock up to 2 miles wide. Evidently a similar situation exists in Tobago, for the dikes intruded into volcanics, north of a line between Rocky Point and the north edge of Providence village, dip toward the batholith and form a part of the fan pattern. It will be shown later that dikes south of this line belong to systems related to the volcanics.

Most of the faulting observed in the diorite is associated with intense shearing and seems to be due to post-intrusive stresses. Nevertheless, relatively low angle normal faulting, possibly related to the period of intrusion, was observed in two localities. One of these is in the contact zone of the south flank of the batholith, $1\frac{3}{4}$ miles southeast of Plymouth, where, in two parallel faults 150 feet apart, contact-metamorphosed volcanics are down-dropped northward against inclusion-rich meladiorite. At the second locality, on the east side of Culloden Bay, a normal fault in marginal diorite strikes N 55° E and is downthrown to the southeast. Balk (1937, p. 105) regards such faults as an expression of horizontal expansion of the hood of the batholith due to continued injection from below, and thus the tension joints and normal faults associated with the Tobago batholith are probably expressions of the same intrusive forces.

Intrusive agglomerate in the vicinity of Bacolet Point (Fig. 2; Pl. 1) is associated with a dike fan similar to that observed in the diorite. Relatively thin andesite dikes dip inward toward the agglomerate body which outcrops on the coast a mile southeast of Scarborough. Three near-by intrusive agglomerate bodies may belong to this same eruptive center, the absence of dikes away from the coast being attributed to the scarcity of outcrops. The associated fan of dikes probably represents a cone sheet type of intrusion, such as those accompanying centers of volcanic activity in the Mull district (Bailey, *et al.*, 1924; Richey, *et al.*, 1930) and in the Sunlight volcanic area of Wyoming (Parsons, 1939). According to Anderson (in Bailey, *et al.*, 1924, p. 11), cone sheet dikes fill tension fractures produced by pressure of intrusion from below, and are thus somewhat analogous to the fan of tension joints commonly developed in and adjacent to batholiths and stocks.

A group of dikes, apparently unrelated to those of Bacolet Point, outcrops in the Minister Point area (Pl. 1). These have vertical or very steep dips. On the west side of Minister Point a 10-foot thick dike of andesite porphyry strikes northward at its northern end, but curves across the shallow waters of the bay and disappears into

the sea cliff at its southern end with a strike of N 50° W. Other dikes in the vicinity vary in strike in such a way as to enclose an elliptical area centering near the east side of Minister Point. The presence of intrusive agglomerate on Minister Point suggests that this area was also a center of intrusion, with the dikes representing the innermost ring of a cone-sheet system.

Between the dike systems associated with the batholith and with the Bacolet intrusive center a few dikes occur which seem to belong to neither system. Their strikes suggest one or more separate centers of intrusion. No bodies of intrusive agglomerate were found in the area, however, and the number of dikes mapped is too small to permit any definite conclusions as to the location of the center or centers.

Quartz-epidote and quartz veins are helpful in distinguishing fractures related to intrusive forces from those caused by later regional deformation. The veins occur in schist, diorite, and volcanics (except the Hawk's Bill group), and are especially abundant in the marginal facies of the diorite, in volcanics and schist near the diorite contact, and in volcanics near the Bacolet and Minister Point intrusive centers. Hence the vein-forming solutions probably emanated from the diorite and from deep-seated intrusions in the Bacolet area. Wherever quartz veins and andesite dikes occur together, they tend to be parallel, but the veins are younger because in some places they cut the dikes. The evidence indicates, therefore, that the veins are younger than much of the diorite, are at least slightly younger than the accompanying andesite dikes, are older than the Hawk's Bill volcanics, and are thus related to a late but not the latest stage of igneous activity.

Quartz veins are particularly abundant in the fine tuffs of the Bacolet area. They average less than an inch thick and occur in a rhomboidal pattern suggestive of joint control. Some are displaced an inch or two by small faults, the fault planes themselves being occupied by quartz veins. In the same area quartz-epidote veins are most common in crush zones associated with minor faulting; as for example, the normal fault on the east side of Bacolet Point (Fig. 2), where a quartz-epidote vein occupies the fault plane. These vein-filled faults and fractures are thus probably related to intrusive stresses.

Normal faults occur on either side of the Bacolet intrusive center, downthrown away from the center. One of these faults contains the quartz-epidote vein mentioned above. Probably all were produced in the intrusive period that culminated in the emplacement of the cone-sheet dikes, both dikes and fractures being an expression of the vertical movements involved.

*Structures related to regional deformation.*—Regional deformation produced intense shearing and small-scale thrust and strike-slip faulting in all rocks older than the late Tertiary sediments, the shearing planes passing uninterruptedly through dikes (Pl. 6, fig. 2) and quartz veins. For the most part, these movements produced crush zones, but little or no chemical alteration in the rocks affected. Locally, however, the volcanics were chloritized along shear zones. An excellent example of chloritization may be seen in the road-metal quarry on the northern outskirts of Scarborough, where fine tuff-breccia, involved in intricate shearing and small-scale thrust-faulting, was converted to a rock resembling a greenschist. The associated quartz-epidote veins were shattered and the fragments rotated by the movements. Quite probably

this is the rock mapped as greenschist by Cunningham-Craig (1907), for no other rocks resembling schist were seen in the Scarborough area.

During regional deformation it is possible that the volcanics yielded to some extent by folding as well as by shearing. Wherever bedding can be detected, the pyroclastic rocks are moderately to steeply tilted. In the Bacolet area, for example, dips vary from 10° to 45°, while near Goldsborough Bay dips of 55° were observed. Nearly all of the beds dip to the northwest however. There is little evidence of reversal except in the fine tuffs near Mount St. George, and here the reversals are sharp and apparently connected with faulting. The coarse pyroclastics associated with intrusive centers on the south coast do not seem to be involved in strong folding, for they dip northward under finer tuffs and do not again appear at the surface. The distribution of coarse and fine debris is, in fact, exactly that which would be expected in undisturbed pyroclastics, and the presence of vertical agglomerate and breccia bodies cutting the dipping pyroclastics gives further argument against any great amount of folding. Apparently the beds owe their present attitudes partly to high initial dips and partly to a steepening of initial dips during deformation. The amount of steepening would not have to be great. Slopes of 15° to 20° are common in pyroclastic debris around modern volcanic cones. Hovey (1902, p. 340) records slopes of 30° in the upper parts of the old cone of Mont Pelée, while Kuenen (1935, p. 63) reports subaerial slopes of 13° to 37° on some young East Indian volcanic cones.

Thrusting and shearing in diorite and volcanics of the western part of Tobago indicate a shortening of the section in a generally north-south direction. Likewise, the east-west folds and northward overturning visible near the west end of the schist outcrop point to a similar direction for the pre-diorite schist-forming movements, accompanied by underthrusting from the north or overthrusting from the south. Toward the east end of the island, the average strike of minor folds within the schist swings to the northeast, with overturning in a northwesterly direction. The elongation of the diorite mass follows this swing to the northeast. Structural data are insufficient to demonstrate a similar change in the direction of the stresses causing post-intrusive deformation in the diorite and volcanics, but it seems probable that structures in the schist are indicative of the stresses acting during the two or more major periods of deformation which have affected the island.

*Relative age of the diorite and volcanics.*—Evidence bearing on the relative age of the diorite and volcanics is somewhat contradictory. Certain data indicate essential contemporaneity. For example, variations in composition within the diorite approximate the variations found in the rocks of the volcanic group. Furthermore, the dikes cutting both the batholith and the volcanics are identical in composition and seemingly have been emplaced along fractures produced by intusive forces. Thus, a common magma source seems to be indicated for the dikes at least, and possibly for all the igneous material, both intrusive and extrusive. On the other hand, the diorite always bears intrusive relations to the pyroclastic rocks (of pre-Hawk's Bill age) and might be considered distinctly younger. Finally, the hornblende diorite xenolith found in tuff-breccia on Goat Island indicates that some diorite is older than a part of the volcanics.

The writer believes that the diorite was derived from the same magma source as

the extrusive volcanics, that diorite bodies underlie volcanic centers, such as those in the Bacolet area, and that both intrusive rocks and volcanics vary in age from place to place. Evidence indicates a westward shift of volcanic activity on Tobago. The Goldsborough dacitic tuffs and breccias are the oldest and are intruded by coarse-grained, mesocratic diorite, and quartz diorite, probably representing the underlying magma which fed the volcanic vents. The Minister Point volcanic center was apparently next oldest, followed by centers near Bacolet Point. Diorite does not outcrop in the Bacolet area, but the andesite cone-sheet dikes point to its presence at depth. Merchiston volcanics resemble those of the Bacolet formation and may be the same age. If so, the xenolith of diorite in the former could have come from the intrusive body connected with the Goldsborough volcanics. Hawk's Bill volcanics on the west coast are the youngest, younger than the period of formation of quartz-epidote veins but older than at least part of the andesite dikes. These volcanics are, therefore, contemporaneous with a very late stage in the emplacement of the diorite batholith and probably represent the last extrusive activity occurring on Tobago.

## TERTIARY AND QUATENARY SEDIMENTS

### GENERAL STATEMENT

Near the villages of Scarborough and Bethel, patches of undeformed, fossiliferous clays, sands, and marls of late Tertiary age lie unconformably on the Bacolet volcanics. In previous reports (Cunningham-Craig, 1907; Trechman, 1934), these sediments have been referred to as the "Tertiary beds", no formal name having been applied. Inasmuch as the best exposures of these "Tertiary beds" are found along the shores of Rockly Bay southwest of Scarborough, the writer proposes the name "Rockly Bay formation" for all of the Tertiary strata described below, though the marly beds near Bethel might reasonably be considered a separate formation.

Rotted coral limestone of probable Quaternary age covers the low-lying southwest tip of the island, and outliers of coral occur on top of the sea cliff a mile northeast of Culloden Bay and at Courland Point near Plymouth. Clays and sands underlying the coral are correlated provisionally with the Rockly Bay formation.

### ROCKLY BAY FORMATION

*Scarborough area.*—In his description of the Tobago Tertiary beds near Scarborough, Trechman (1934) recognized two facies, the *Arca patricia* bed and the Tobago Crag or *Balanus* clays. The former, exposed on the west side of Red Point 2 miles southwest of Scarborough, have a variable thickness of coarse conglomerate at the base and lie on a weathered and eroded surface of tuff-breccia. The following section is essentially that given by Trechman (1934, p. 483) with the addition at the top of 23½ feet of beds outcropping along a foot-path across the road and above the section measured by Trechman.

*Arca patricia* Beds, Red Point

| *Top* | *Feet* |
|---|---|
| Pebble bed, many specimens of Ostrea; Calcareous cement | 2 |
| Fine sand, loosely cemented by calcite and limonite. Unfossiliferous | 1½ |
| Gravel in path, no fresh section | 15 |

| | |
|---|---|
| Covered. Probably soft, brown clay | 5 |
| Soft, brown clay with calcareous concretions; a few *Arca patricia* shells | 6 |
| Pebble bed with a small variety of *Ostrea* | 3 |
| Soft, brown, sandy clay with specimens of large *Ostrea* | 3 |
| Hard, brown, sandy clay with scattered pebbles and abundant *Ostrea* and *Arca patricia* shells | 2 |
| Layer rich in shells of a small *Ostrea* | 1 |
| Conglomeratic layer carrying *Arca patricia* shells | 2 |
| Pebble bed | 2 |
| Bed carrying many specimens of small *Ostrea* | 2 |
| Coarse conglomerate, angular to subangular pebbles and cobbles, chiefly basalt porphyry from underlying volcanics. Some pebbles of quartz-epidote and quartz vein material | 4–7 |
| Volcanics, somewhat weathered, about at sea level | — |

Trechman described greenschist pebbles from the basal conglomerate as resembling the rocks (schists) exposed on the north and west side of Tobago. The observation is of interest, for it indicates a possible occurrence of schist in the Scarborough area. However, the writer was unable to find schist pebbles here or in other Tertiary conglomerates in the vicinity of Scarborough. The possibility that the sheared and chloritized tuff-breccia exposed north of Scarborough might be mistaken for chlorite schist has been mentioned previously.

The Tobago Crag or *Balanus* clay beds are well exposed in a sea cliff 1¼ miles southwest of Scarborough. Because of the lenticularity of these beds, the following section differs somewhat from that measured by Trechman in the same area.

*Balanus* Clay Beds

| *Top* | *Feet* |
|---|---|
| Brown soil | 1½ |
| Fine white sand, unfossiliferous | 2 |
| Thin-bedded gray and brown clay and sandy clay. No fossils | 13 |
| *Balanus* beds. About 75% *Balanus* fragments in matrix of brown, sticky clay. A few pebbly layers. Some specimens of *Ostrea*, *Pecten* and a few other mollusks. Bed seems to be a channel filling. Passes laterally into gray clay | 18 |
| Brown clay with calcareous nodules. Specimens of *Balanus*, *Ostrea* and a few other mollusks, at beach level | 1 |

At the Government Stock Farm on the west side of Lambeau river valley, beds consisting of 80% *Balanus* fragments in a light-brown clay matrix are exposed up to an elevation of 150 feet, with tuff-breccia outcropping above this level. Fragments of *Balanus* are found on the hillside below the stock buildings, almost to the bank of Lambeau river, which has cut a 6-foot channel into brown clay and pebble layers similar in appearance to Rockly Bay beds exposed along the coast. On the east side of the Lambeau river valley Tertiary sediments are found as high as 50 feet above sea level. Apparently the *Balanus* beds were deposited to a depth of at least 150 feet in a Miocene forebearer of the Lambeau river valley and the river subsequently re-excavated its channel, leaving only remnants of the Tertiary deposits along its valley walls.

North and northwest of Scarborough, weathered yellow and reddish-brown sandy clays and pebble beds are exposed intermittently in road cuts up to an elevation of

about 190 feet. No fossils were found in these beds, though, on the Scarborough-Plymouth road, a conglomerate carrying fragments of *Ostrea* lies at the base of the unfossiliferous beds. These conglomeratic layers are believed to be the basal remnants of the Rockly Bay formation, now largely removed by erosion.

Beds which possibly are to be correlated with the *Balanus* clays outcrop in the Bacolet river valley, 1½ miles east of Scarborough. A 25-foot section of fossiliferous brown clays bearing large calcareous nodules is exposed along Windward road on the west side of the valley. Igneous material outcrops on top of Bacolet Point above the fossiliferous beds, and the latter lie on weathered tuffs at an elevation of 30 feet, near the base of the road cut. Fossil-bearing float can be followed southward for a distance of 300 yards along the east side of Bacolet Point, but the beds do not outcrop at the coast. About a quarter mile to the north, along Friendsfield road, a 2-foot thick bed of lime-cemented conglomerate, carrying *Arca patricia*, forms a narrow bench along the valley wall at an elevation of 45 feet. Igneous rock is exposed above and below the conglomerate. No Tertiary sediments were found on the east side of Bacolet river valley. As in the Government farm area, these beds are believed to have been deposited in a valley of much the same location and depth as that now occupied by the present river, which has since re-excavated its channel.

Dr. H. G. Richards of the Philadelphia Academy of Science kindly identified the following fossils collected by the writer from the beds of the Scarborough area:

From section of Balanus beds 1¼ miles southwest of Scarborough:

*Glycimeris* cf. *aculicostatus* Moore
*Vermetus* sp.
*Balanus tintinabulum* linnaeus

From conglomerate bed, west side of Bacolet river valley:

*Arca patricia* Sowerby

These forms were also described from the Scarborough area by Dr. Trechman, who considered the beds to be of late Miocene age. On the basis of the small amount of material available to him, Dr. Richards reached the same conclusion.

From a study of the small foraminifera, H. H. Renz (personal communication) believes the Balanus beds can be compared with the lower part of the Talparo formation of Trinidad. This formation is of latest Miocene and Pliocene age. Dr. Renz has found reworked Eocene Orbitoids in the base of the Tobago *Balanus* beds, a rather surprising occurrence, since fossiliferous Eocene strata are not otherwise represented on the island.

*Bethel area.*—West and southwest of Bethel, white to light-brown chalky deposits are exposed in road cuts high above the surrounding coral. Along the Old Grange road a mile southwest of Bethel, coral occurs as high as 100 feet above sea level. Between 100 and 150 foot elevations, weathered volcanic tuffs occur in the road cuts. From 150 to 185 feet the chalky deposits are exposed, and above 185 feet tuffs outcrop once more. The chalky material contains broken fragments of *Balanus*, *Ostrea*, and other mollusks, as well as weathered pebbles of volcanics. A similar deposit is found west of Bethel, on the road leading to Mount Irvine Bay, between 210 and 240 feet above sea level. Trechman (1934, p. 490) listed a few fossils from the exposure along Old Grange road and assigned the beds questionably to the Pliocene.

*Tertiary (?) sediments underlying coral.*—In 1911 and 1912 the Tobago Oil Syndicate, Ltd. drilled two exploratory wells through the coral at the southwest end of the island.[3] Well No. 1, "Lowlands Estate", is located on the southernmost part of Tobago within a few hundred feet of the shore (Pl. 1), and well No. 2, "Cove Estate", lies a mile to the northward. Since neither well reached basement rocks, the total thickness of Cenozoic sediments at the south end of Tobago is not known, but it exceeds 500 feet.

No. 1
"Lowlands Estate"

| *Feet* | Elevation about 10 feet |
|---|---|
| 0– 21 | Loose coral limestone snags with red soil |
| – 40 | Coral limestone |
| – 60 | Whitish clay |
| – 72 | Bluish clay |
| –175 | Blue clay (darker) |
| –180 | Gray quicksand |
| –282 | Blue clay (dark) |
| –295 | Gray quicksand with lignite |
| –357 | Blue clay (dark) |
| –420 | Gray quicksand |
| –509 | Dark-blue clay |

No. 2
"Cove Estate"

| *Feet* | Elevation about 20 feet |
|---|---|
| 0– 80 | Yellow clay |
| –138 | Blue clay |
| –142 | Gray quicksand |
| –255 | Blue clay |
| –263 | Coarse gray grit with pieces of shell |
| –296 | Very fine dark-gray grit |
| –488 | Blue-gray gravel |

Correlation of the two logs is rather difficult; if the thin beds of gray quicksand found at a depth of 180 feet in well No. 1 and 142 feet in well No. 2 are equivalent, a southerly dip of about 50 feet per mile between the two wells is indicated. So far as the writer knows, there is no paleontologic evidence as to the age of the beds beneath the coral. The preponderance of clay in the well logs suggests a correlation with the *Balanus* clays of the Rockly Bay formation.

### QUATERNARY BEDS

*Description.*—Coral limestone covers the low-lying southwest tip of Tobago, rising gently from a low bench at the sea coast to an elevation of about 100 feet and overlapping on volcanic tuffs to the northward. Outliers of coral lie on quartz diorite at Courland Point and on meladiorite a mile northeast of Culloden Bay. At the latter locality, conglomerate bearing coral and mollusk fragments occurs on top of the sea cliff, approximately 150 feet above sea level.

The coral rock, everywhere porous and recrystallized, consists mostly of broken fragments of coral and mollusks, with branching forms of coral predominating. An exposure on the south side of Mount Irvine Bay is typical of the coral as it appears along the coast. The section follows:

| *Top* | *Feet* |
|---|---|
| Rotted coral, mostly branching forms | 13 |
| Poorly bedded, fine white sand, somewhat iron-stained. Small calcareous nodules | 6 |
| Rubble of rotted coral rock, to beach level | 5 |

[3] The writer is indebted to Drs. H. H. Suter and H. G. Kugler of Trinidad Leaseholds Ltd. for information on these wells.

about 190 feet. No fossils were found in these beds, though, on the Scarborough-Plymouth road, a conglomerate carrying fragments of *Ostrea* lies at the base of the unfossiliferous beds. These conglomeratic layers are believed to be the basal remnants of the Rockly Bay formation, now largely removed by erosion.

Beds which possibly are to be correlated with the *Balanus* clays outcrop in the Bacolet river valley, 1½ miles east of Scarborough. A 25-foot section of fossiliferous brown clays bearing large calcareous nodules is exposed along Windward road on the west side of the valley. Igneous material outcrops on top of Bacolet Point above the fossiliferous beds, and the latter lie on weathered tuffs at an elevation of 30 feet, near the base of the road cut. Fossil-bearing float can be followed southward for a distance of 300 yards along the east side of Bacolet Point, but the beds do not outcrop at the coast. About a quarter mile to the north, along Friendsfield road, a 2-foot thick bed of lime-cemented conglomerate, carrying *Arca patricia*, forms a narrow bench along the valley wall at an elevation of 45 feet. Igneous rock is exposed above and below the conglomerate. No Tertiary sediments were found on the east side of Bacolet river valley. As in the Government farm area, these beds are believed to have been deposited in a valley of much the same location and depth as that now occupied by the present river, which has since re-excavated its channel.

Dr. H. G. Richards of the Philadelphia Academy of Science kindly identified the following fossils collected by the writer from the beds of the Scarborough area:

From section of Balanus beds 1¼ miles southwest of Scarborough:

*Glycimeris* cf. *aculicostatus* Moore
*Vermetus* sp.
*Balanus tintinabulum* linnaeus

From conglomerate bed, west side of Bacolet river valley:

*Arca patricia* Sowerby

These forms were also described from the Scarborough area by Dr. Trechman, who considered the beds to be of late Miocene age. On the basis of the small amount of material available to him, Dr. Richards reached the same conclusion.

From a study of the small foraminifera, H. H. Renz (personal communication) believes the Balanus beds can be compared with the lower part of the Talparo formation of Trinidad. This formation is of latest Miocene and Pliocene age. Dr. Renz has found reworked Eocene Orbitoids in the base of the Tobago *Balanus* beds, a rather surprising occurrence, since fossiliferous Eocene strata are not otherwise represented on the island.

*Bethel area.*—West and southwest of Bethel, white to light-brown chalky deposits are exposed in road cuts high above the surrounding coral. Along the Old Grange road a mile southwest of Bethel, coral occurs as high as 100 feet above sea level. Between 100 and 150 foot elevations, weathered volcanic tuffs occur in the road cuts. From 150 to 185 feet the chalky deposits are exposed, and above 185 feet tuffs outcrop once more. The chalky material contains broken fragments of *Balanus*, *Ostrea*, and other mollusks, as well as weathered pebbles of volcanics. A similar deposit is found west of Bethel, on the road leading to Mount Irvine Bay, between 210 and 240 feet above sea level. Trechman (1934, p. 490) listed a few fossils from the exposure along Old Grange road and assigned the beds questionably to the Pliocene.

*Tertiary (?) sediments underlying coral.*—In 1911 and 1912 the Tobago Oil Syndicate, Ltd. drilled two exploratory wells through the coral at the southwest end of the island.[3] Well No. 1, "Lowlands Estate", is located on the southernmost part of Tobago within a few hundred feet of the shore (Pl. 1), and well No. 2, "Cove Estate", lies a mile to the northward. Since neither well reached basement rocks, the total thickness of Cenozoic sediments at the south end of Tobago is not known, but it exceeds 500 feet.

No. 1
"Lowlands Estate"
Elevation about 10 feet

| Feet | |
|---|---|
| 0- 21 | Loose coral limestone snags with red soil |
| - 40 | Coral limestone |
| - 60 | Whitish clay |
| - 72 | Bluish clay |
| -175 | Blue clay (darker) |
| -180 | Gray quicksand |
| -282 | Blue clay (dark) |
| -295 | Gray quicksand with lignite |
| -357 | Blue clay (dark) |
| -420 | Gray quicksand |
| -509 | Dark-blue clay |

No. 2
"Cove Estate"
Elevation about 20 feet

| Feet | |
|---|---|
| 0- 80 | Yellow clay |
| -138 | Blue clay |
| -142 | Gray quicksand |
| -255 | Blue clay |
| -263 | Coarse gray grit with pieces of shell |
| -296 | Very fine dark-gray grit |
| -488 | Blue-gray gravel |

Correlation of the two logs is rather difficult; if the thin beds of gray quicksand found at a depth of 180 feet in well No. 1 and 142 feet in well No. 2 are equivalent, a southerly dip of about 50 feet per mile between the two wells is indicated. So far as the writer knows, there is no paleontologic evidence as to the age of the beds beneath the coral. The preponderance of clay in the well logs suggests a correlation with the *Balanus* clays of the Rockly Bay formation.

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Near the middle of Mount Irvine Bay the coral overlies brown, pebbly clay and conglomerate. No fossils were found in the clay, but in appearance it resembles the Rockly Bay beds exposed near Scarborough.

Well No. 1, "Lowlands Estate", at the extreme south edge of the island, penetrated 40 feet of coral before entering clay. The 80-foot thickness of yellow clay logged at the top of well No. 2 probably was derived largely from the weathering of the coral. For the southwestern tip of the island, then, a thickness of 40 to 80 feet of coral is indicated. Probably it was never much thicker, for the surface has been but little dissected by stream erosion.

Except for the outlier northeast of Culloden Bay, coral is not found at elevations greater than 100 feet above sea level. It is quite possible that this outlier is to be correlated with the Rockly Bay beds rather than with the Quaternary coral. There is no evidence that the coral ever covered a much larger proportion of Tobago than it does at present.

Along Providence river, 3 miles north of Scarborough, a small patch of sandy clay and pebbly beds occurs well down within a valley cut in volcanic tuffs. The conglomerate layers contain only porphyry and quartz-epidote pebbles. These beds are 30 feet thick, and their top lies about 110 feet above sea level. Most probably they represent a channel filling, deposited in Providence river valley when the Quaternary seas approached their highest level.

*Age of the Tobago coral.*—Trechman (1934, p. 492) identified a number of fossils from the base of the coral near Plymouth. He stated that the forms are of recent aspect. In comparing the Tobago coral with that of Barbados, which may be partly Pliocene, he concluded the Tobago coral". . . to be a formation of later commencement and lesser duration than that of Barbados." It would seem, therefore, that most of the Tobago coral is to be referred to the Quaternary period, though the beds at higher elevations could be of Pliocene age.

## SUMMARY OF TERTIARY AND QUATERNARY HISTORY

Very little is known concerning the early Tertiary geologic history of Tobago. Apparently the area was above sea level during the extrusion of the Upper Cretaceous (?) volcanics, for no marine sediments are found in the volcanic group. The presence of reworked Eocene fossils in the base of the *Balanus* beds, however, suggests a partial submergence of the island during the Eocene. Thereafter, it apparently remained above sea level, developing essentially its present stream pattern before latest Miocene time. During a period of submergence, which probably extended into the Pliocene, the rocks of the Miocene sea coast were covered with coarse conglomerate and sandy clay, and the drowned river valleys were filled with mud and sand. At this time the sea apparently reached a height of at least 240 feet above the present level.[4]

The island probably emerged in Pliocene time, allowing the removal of much of the Tertiary sediments before the formation of the Quaternary coral reefs. During the Quaternary submergence the sea rose to little more than 100 feet above the present

[4] At this height less than 15% of the island's present area would be submerged.

level. Unconsolidated sands and gravels found in Providence river valley probably owe their existence to drowning of the valley when the Quaternary seas stood highest. Subsequent emergence resulted in re-excavation of the drowned valleys and removal of most of the debris which had filled them. The valleys may have been cut somewhat below their present depths, for the embayed and cliffed coastline of Tobago would seem to indicate recent submergence, or more probably a rise of sea level.

## GEOLOGIC HISTORY OF TOBAGO

### REGIONAL SETTING

*Introduction.*—Although the ages of the Tobagan formations can be determined relative to each other, only the Pleistocene coral and fossiliferous Tertiary beds can be definitely correlated with rocks of other areas. Nevertheless, certain probable correlations are suggested by the descriptions of similar sequences of rocks and related diastrophic movements elsewhere in the Caribbean. For the most part, only those references are cited which are pertinent to this interpretation. For a more complete bibliography see Schuchert (1935) or L. Rutten (1938).

*Greater Antilles and Virgin Islands.*—Except for Upper Jurassic marine sediments in Cuba (Palmer, 1945; De Albear, 1947; Vermunt, 1937), the oldest rocks of this area which can be dated are Cretaceous. Volcanic tuff, tuff-breccia, and lava flows make up a large proportion of the Cretaceous deposits, with andesitic types predominating. Ultramafic rocks intrude Turonian or younger volcanics in western and central Cuba (M. G. Rutten, 1936, p. 4; Vermunt, 1937, p. 18). Keijzer (1945, p. 13) assigns a similar age to serpentines of eastern Cuba, but Thayer and Guild (1947) show that the ultramafics of this area are older than at least part of the Upper Cretaceous tuffs. Ultramafics intrude Upper Cretacous volcanics in Puerto Rico (Hubbard, 1923, p. 38; Meyerhoff, 1933, p. 135), and Cretaceous sediments in Hispaniola (Weyl, 1941, p. 10). Small and large masses of diorite, quartz diorite and granodiorite were emplaced later. In Cuba latest Cretaceous beds, Maestrichtian and probably upper Senonian, rest unconformably on older Cretaceous volcanics, intrusive diorites, and ultramafics. Hence, if the ultramafics are of a single generation they are probably of late Cretaceous, about early Senonian age.[5] Although they have not been closely dated on Hispaniola and Puerto Rico, there seems to be no evidence against a similar age for the ultramafic rocks of these two islands.

According to Palmer (1945, p. 27), the earliest and strongest deformation in Cuba took place within the Upper Cretaceous. The faunal study by Keijzer (1945, p. 151–154) suggests the orogeny occurred in the early Senonian. In Hispaniola an apparently equivalent period of deformation is dated as Laramide by Weyl (1941, p. 28). Because of the apparent absence of post-orogenic Cretaceous and Eocene beds in Puerto Rico the comparable orogeny, if it occurred, cannot be closely dated.

A second great orogenic period in the late middle Eocene (Palmer, 1945, p. 30) resulted in intense overthrusting and folding in eastern and central Cuba (Thayer and Guild 1947, p. 926; M. G. Rutten, 1936, p. 5). This period of deformation

[5] Palmer (1945, p. 18) stated that serpentine masses of several different ages are present in Cuba, though "most of them are intrusive within the Upper Cretaceous Terrain."

probably affected the remainder of the northern islands but has been recognized only in St. Croix, where Maestrichtian beds have been strongly deformed (Imlay, 1944, p. 1007), and possibly in Hispaniola (Weyl, 1941, p. 26). Mild to strong Oligocene folding occurred locally in Cuba (M. G. Rutten, 1936, p. 28) and in Hispaniola (Vaughan *et al.*, 1922, p. 87). In the Miocene and early Pliocene, Hispaniola suffered strong folding and faulting, Cuba experienced a mild orogeny (Palmer 1945, p. 32) and the rocks of St. Croix were gently folded and strongly faulted (Cederstrom, 1941, p. 576).

*Lesser Antilles.*—The Lesser Antilles comprise two groups of islands, an inner (western) group or arc, marked by recent volcanism, and an outer (eastern) group, the "limestone Caribbees," characterized by ancient volcanic rocks and interbedded sediments, overlain by younger limestone. To this latter group belong Sombrero, Anguilla, St. Martin, St. Barthelemy, Barbuda, Antigua, Grande Terre of Guadeloupe, Desirade, and Marie Galante. Both Senn (1940) and Woodring (1928) have summarized the geologic history of this group, which will be reviewed here only briefly.

In the outermost islands of this group, Sombrero and Barbuda, only Quaternary limestone is exposed. Volcanic basement rocks outcrop on Anguilla, St. Martin, and St. Barthelemy. On Anguilla, Oligocene limestone rests unconformably on the basement rocks and Eocene or Oligocene limestone covers the St. Martin volcanics. The basement rocks of St. Martin consist of well stratified, strongly folded and metamorphosed tuff, tuff-breccia, and crystalline limestone, intruded by a quartz diorite-pyroxene diorite complex. According to G. A. F. Molengraaff (1931, p. 735), the sedimentary material of the basement may be of Cretaceous age, comparable to similar rocks on St. Thomas. Lower Tertiary limestone overlying the basement rocks has been less deformed than the basement, and Miocene limestones are perfectly horizontal. Two periods of folding are indicated, one pre-early Tertiary and the other pre-Miocene.

The basement rocks of St. Barthelemy consist of volcanic debris intruded by a "batholith" which Cleve (1871) identified as syenite porphyry. Samples from the margins of the igneous body, furnished the writer by Dr. Alfred Senn, are dacite porphyry or fine-grained augite quartz diorite, suggesting that the body is a quartz-bearing diorite, as are similar intrusions throughout the Antilles. A thick series of volcanic tuffs and agglomerates, with interstratified upper Eocene limestone beds, rests unconformably on the basement rocks on St. Barthelemy (Senn, 1940, p. 1592).

Both Antigua and Desirade have volcanic basement rocks. In Antigua, a great thickness of andesites, tuffs, and agglomerates of probable Eocene age is overlain in turn by lower Oligocene limestone and fine tuffs and by upper Oligocene limestone. Cretaceous chalk, discovered while deepening a water well in the Oligocene tuffs, is regarded by Senn (1940, p. 1599) as probably representing the pre-volcanic basement, though other evidence strongly indicates the chalk is not in place nor of local origin (Trechman, 1941, p. 117). Desirade possesses a basement of andesite and rhyolite flows and intrusive granodiorite (Barrabe, 1934). Miocene limestone unconformably overlies the basement.

Grande Terre and Marie Galante are the southernmost, and also the innermost of

the "limestone Caribees". The latter is covered by a cap of recent limestone (Woodring, 1928, p. 417). Grande Terre, however, has a basement of granodiorite, overlain unconformably by Miocene tuff and limestone (Barrabe, 1934).

To summarize, the "limestone Caribbees" are characterized by andesitic lava flows and coarse volcanic debris of Eocene age or older. Plutonic dioritic rocks, some or all pre-late Eocene, intrude the volcanic basement. Oligocene and younger beds are mostly limestones, and volcanic debris, where present, is fine-grained (as the Central Plain Tuff of Antigua). Apparently these islands were active volcanic centers in pre-Oligocene time and have since received volcanic debris only sporadically and from a distance. The islands have undergone at least two periods of deformation, one pre-early Eocene and one between early Eocene and Miocene time.

The inner arc of the Lesser Antilles, stretching from Saba to Grenada, is characterized by recent or sub-recent volcanic activity. Tuffs of Oligocene age on Martinique (Senn, 1940, p. 1596) and of Oligocene or early Miocene age on Carriacou (Trechman, 1935b, p. 553) represent the oldest beds identified in the inner arc. Volcanic activity started here at least as early as Oligocene and continued with few interruptions to the present. As in the older (pre-Oligocene) volcanics of the Greater and Lesser Antilles, andesites predominate, with basalts and dacites also present. However, the andesites and basalts of the more recent volcanics contain hypersthene as a common constituent while, so far as the writer could determine, this mineral is uncommon in the pre-Oligocene volcanics. The significance of this mineralogic variation is not apparent.

*Margarita and the Dutch Leeward Islands.*—The northern part of Margarita[6] is composed of ortho- and paraschists and gneisses, intruded by serpentinized peridotite and dunite and by small bodies of diorite or gabbro.[7] A zone of slightly metamorphosed clastic sediments with thin limestone beds lies south of the schist area. Unmetamorphosed sandstones and shales with coraliferous limestone at the base, all apparently of Eocene age, are in unconformable contact with low grade metamorphics and are folded into a syncline along the south coast. Conglomerates near the base of the Eocene section contain numerous volcanic pebbles and cobbles, chiefly andesites. Miocene sediments lie unconformably on the lower Tertiary strata.

In Margarita, as in Tobago, the foliation in the schist strikes slightly north of east and dips steeply southeast. Locally the peridotite shows hydrothermal alteration, possibly related to the "diorite" intrusions, though this is the only evidence bearing on the relative age of the two. As in Tobago, the "diorite" has suffered strong shearing. The major deformation in Margarita predated at least part of the Eocene; it may be dated more precisely with the completion of current faunal studies. Detrital grains of chromite and enstatite in Eocene sands prove that the ultramafic mass had been exposed to erosion by that time. Sometime between the Eocene and late Miocene, the lower Tertiary sediments were folded into a syncline with its axis approximately parallel to the foliation in the schist.

[6] From unpublished data collected by H. H. Hess and others, including the writer during, visits to the island in 1939 and 1947.

[7] The diorite or gabbro is coarse grained, mesocratic, and resembles a diorite, but augite is more abundant than hornblende in the specimens collected. The feldspar is completely saussuritized and could not be identified.

The Dutch Leeward Islands, Aruba, Curacao, and Bonaire, are volcanic in character, comparable to the Greater Antilles and the outer islands of the Lesser Antilles. A deformed basement of intrusive and extrusive gabbroic and dioritic rocks with intercalated radiolarian cherts, limestone, and graywack is present on all three islands. Quartz-augite diorite in Curacao (G. J. H. Molengraaff, 1931) and quartz diorite in Aruba (Westermann, 1931) intrude the basement rocks. On Bonaire, limestone of probable latest Cretaceous age unconformably overlies the basement rocks, which likewise are believed to belong to the Upper Cretaceous (Pijpers, 1933). On Curacao, a series of coarse detrital sediments overlies the limestone and is folded with it. Upper Eocene limestone is not involved in the folding. In the Dutch Leeward Islands then, a basement of volcanic and intrusive rocks was deformed in pre-latest Cretaceous time, unconformably overlain by Upper Cretaceous-Eocene (?) sediments, and again folded prior to late Eocene.

*Barbados, Trinidad and Northeastern Venezuela.*—The geology of Barbados has been discussed in detail by Senn (1940). Clastics of early and middle Eocene age are the oldest beds exposed. These beds were uplifted, strongly folded and thrust-faulted, and eroded, then covered by a thick series of mud flows. Upon the strongly folded clastic sediments and mud flows were deposited the Oceanic beds, a considerable thickness of upper Eocene chalk, radiolarian earth, and tuff. Deformation apparently continued in Barbados during the deposition of the Oceanics, for these beds are also folded and faulted, though much less so than the older formations. The Oceanic beds, in turn, were uplifted and eroded, then submerged and covered by Upper Oligocene-Miocene marls. There is evidence of Miocene-Pliocene folding and post-Pleistocene uplift and fracturing.

Jurassic and Cretaceous rocks outcrop in Trinidad, supplying evidence as to the pre-Eocene history of the southern West Indian region. Jurassic rocks are found only in the North Range. According to Senn (1940, p. 1562) they consist mainly of phyllites with abundant lenses and veins of quartz and interbedded crystalline limestone. Late Tithonian ammonites have been found in the limestone. Presumably these rocks are equivalent to the Lower Caribbean series of Waring (1926), which he described as calcareous and carbonaceous schists and quartzitic grits. Associated with the Jurassic rocks is a younger system of less metamorphosed dark limestones, grits, and slightly metamorphosed shales, from which Trechman (1935a) collected fossils of late Cretaceous age. The North Range schists have been tightly folded, in general showing axial plane foliation (Waring, 1926, p. 116). They strike a few degrees north of east and are overturned toward the north.

In a small area near the village of Sans Souci, igneous rock identified as "granophyr" intrudes dark calcareous schists of the Lower Caribbean series (Waring, 1926, p. 31). Both massive and pyroclastic igneous rocks are present and on Manantial hill schist seems to be infolded into the igneous mass. Dr. J. C. Griffiths kindly provided a specimen of the intrusive rock from the Sans Souci area. The rock is a fine-grained, holocrystalline augite andesite with a diabasic texture and abundant secondary (probably deuteric) chlorite, leucoxene, and pyrite. It is similar to a specimen which Waring describes from the same area. A fine-grained porphyritic rock consisting principally of andesine, orthoclase and quartz occurs near by.

The North Range seems to have been involved in at least two periods of folding, one in post-late Jurassic and one in post-late Cretaceous time. The igneous rocks at Sans Souci were intruded and extruded between the periods of deformation, though apparently there was relatively little igneous activity in Trinidad. The late Cretaceous sediments probably were folded during the strong basal upper Eocene deformation. An upper Miocene period of folding is also recognized. The North Range schists of Trinidad resemble the Tobago North Coast schists in degree of metamorphism, in the fact that both series are isoclinally folded and overturned northward, and in that both have associated younger andesitic volcanics. On the other hand, Tobago lacks the limestone, carbonaceous schist, and coarse grit of the North Range, and Trinidad has no counterpart of the metavolcanics which comprise the major part of the Tobago schists. The small amount of igneous activity in Trinidad is likewise in marked contrast to the predominantly igneous character of Tobago.

Schists identical with those of the North Range of Trinidad appear in the Serrania de la Costa Oriental of Venezuela (Parkinson, 1925). East of Caracas Cretaceous or, more probably, Jurassic fossils were found in slightly metamorphosed limestone of the coast range (Wolcott, 1943). According to Bucher (1947, p. 108), the first great deformation in northern Venezuela occurred in the Cretaceous, probably in pre-Senonian or intra-Senonian time. Others believe that the Upper Cretaceous orogeny has not been demonstrated and that the major deformation occurred in the late Eocene. During the Miocene-Pliocene deformation, the Serrania del Interior was formed, involving Cretaceous and Tertiary beds in tight folding and southward overthrusting (Hedberg, 1937).

A belt of serpentinite intrusives occurs in the Serrania de la Costa near Caracas, and several peridotite bodies intrude sedimentary rocks of probable Cretaceous age south of the Serrania del Interior. Hence the ultramafics are presumably Cretaceous or younger.

*Summary.*—The diastrophic history of the West Indies began with a great outpouring of andesitic volcanics during the Cretaceous. In late Cretaceous time, but preceding the Maestrichtian, intense deformation occurred, accompanied by the intrusion of ultramafic rocks. Following the emplacement of the ultramafics and before Maestrichtian time, plutonic masses of dioritic composition were likewise intruded. Strong andesitic volcanism continued during the Eocene in the "limestone Caribbees" and in parts of the Greater Antilles.

There was a second major diastrophism near the end of the middle Eocene. Approximately in Oligocene time the inner volcanic arc of the Lesser Antilles became active and volcanism ceased in the outer or "limestone Caribbees". Mild Oligocene deformation occurred locally in Cuba and Hispaniola, but is not recorded for the rest of the West Indian region.

In late Miocene and early Pliocene came the third and last great diastrophism, resulting in rather strong deformation in the Greater Antilles and Venezuela. Uplift and faulting occurred widely during the Quaternary, and modern earthquake shocks give evidence of continuing crustal movement.

## CONCLUSIONS

From the geologic history of the West Indies certain deductions can be made concerning the pre-Miocene history of Tobago. Schists of the North Coast group, predominantely metavolcanic, would seem to correlate with the Cretaceous volcanics which form the basement in most of the Greater and Lesser Antilles, and probably also with the Cretaceous part of the Caribbean series of Trinidad and northeastern Venezuela. The bulk of the debris forming the Tobago North Coast group was derived from near-by volcanoes, while the sediments of the Caribbean series presumably came from the shield area to the south. Quite probably the sericite schist and fine quartzite of the North Coast group likewise represent sediments from the south, carried farther to seaward than the coarser material comprising much of the Caribbean series.

Presumably in late Cretaceous time sediments and volcanics of the North Coast group were strongly folded and metamorphosed. The ultramafic rocks of Tobago were intruded toward the end of the period of intense deformation, and an outbreak of andesitic volcanism followed shortly thereafter. Igneous rocks in the Trinidad North Range may be the southernmost expression of this volcanism. In Tobago, igneous activity ended with the intrusion of the diorite batholith and accompanying andesite dikes. Quite probably the batholith was emplaced in pre-Maestrichtian time, as in Cuba, though the igneous activity may have continued into the Eocene, as in the outer islands of the Lesser Antilles.

The strong middle or late Eocene deformation, recognizable in much of the West Indian area, probably produced the intense shearing of the Tobago volcanics and diorite. During the Eocene the island may have been partly submerged, for reworked orbitoids of this age occur in the *Balanus* beds near Scarborough. Otherwise, there is no positive evidence of submergence, subsequent to the deposition of the North Coast group, until late Miocene time.

The present drainage pattern of the island developed before the end of the Miocene. A partial submergence at this time, continuing into the Pliocene, resulted in the deposition of brackish and fresh water sediments in drowned valleys and along the present coast. These sediments are flat-lying and undeformed. Apparently the Miocene-Pliocene diastrophism did not greatly affect Tobago.

Coral deposits covering the southwest tip of the island were formed during one or more periods of partial submergence in the Quaternary. Relative uplift subsequently resulted in re-excavation of parts of the river valleys which had been filled by sediments. The most recent movement may have been a slight relative submergence, as indicated by the cliffed and embayed shoreline of the island.

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